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THE METEOROLOGICAL BASES OF GLIDER FLIGHT

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CHAPTER 1

GENERAL OBSERVATIONS

A. Conditions Necessary for Glider Flight.

Three factors determine flight performance:

1. Skill and experience.

2. The quality of the glider itself.

3. Knowledge of meteorological conditions favoring flight.

Piloting skill is the basis of all performance. The best glider and strongest upwind cannot correct ineptness. Successful flight, consequently, depends primarily on the pilot. He must be alert, capable and sensitive, in order to judge and apply observations and sensations supplied by nature. This is the essence of motorless flight.

It is possible that a perfectly competent pilot, capable of executing the most intricate maneuvers may be incapable of flying properly a glider if he does not know how to find the energy in the air which constitutes the driving power of the glider.

To observe nature is to love it and to be in communion with it. It is this feeling which enables the pilot to sense the beauty of motorless and noiseless flight -- the most natural method of human flying -- which fills him with enthusiasm and guides him through space in search of the clouds that generate up winds.

The art of piloting, the intelligent assimilation of data furnished by nature and a true feeling for the beauty of glider flight are the distinctive qualities of a pilot capable of accomplishing fine performance.

The aerodynamic qualities of the glider have now attained such perfection that one can hardly hope for much improvement in performances, with the exception of high altitude flights which require stronger and safer construction.

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At the altitudes thus far attained in glider flight nature provides such numerous sources of energy that it is unnecessary to develop irregular and costly structural designs. It seems much more important to concentrate attention on the glider equipment necessary for cloud flying and altitude flying.

The performance of several known gliders is given below:

	"Sperber" Junior	Sao Paulo	"Seeadler" with Floats	"Kranich" Biplane	"Göppingen 3" Minimoa
Wingspread (meters)	16.0	19.0	17.4	18.0	17.0
Surface (sq. meters)	15.5	17.7	18.4	22.7	19.0
Wing load (kg/sq. meter)	18.0	21.6	20.6	20.5	17.5
Speed minimum (km/hr)	48.4	54.0	48.1	47.0	53.7
Vertical speed minimum (meters/sec)	0.74	0.63	0.39	0.75	0.70
Maximum Fineness ratio	24.3	26.0	17.8	22.5	25.7
Vertical speed at 100 km/hr (meters/sec)	1.5	1.36	2.31	1.56	1.52

According to this data, it can be seen that upwind velocities of about 1 meter per second are sufficient to maintain flight. The art of glider flight consists in finding these upwinds.

Since we do not have a distance indicator - the variometer detects only vertical movement at the point where the glider is located - the glider pilot must judge an upwind from meteorological information, cloud form, wind direction and velocity and ground configuration.

Atmospheric phenomena may often appear complex, or even dependent on the laws of chance. However, it should be possible to grasp a clear idea of these changeable phenomena and to formulate an opinion as to what performance is possible in a given situation.

The purpose of this book is to broaden the glider pilot's understanding of atmospheric phenomena. A knowledge of the aerological bases of glider flight should improve the pilot's performance. It should also encourage him to feel that his observations and instrument recordings during flight are contributing to glider progress and research.

Because of their daring new flights and sound judgement of their remarks, many pilots have won the recognition of science and have ensured the lasting value of their performances.

This book is addressed to the practical user and endeavors to afford him, in a simplified and familiar form, an understanding of glider flying.

B. <u>Sources of Energy in Glider Flight</u>.

1. Static and Dynamic Flight.

There are two basic methods of flight: static and dynamic.

Two conditions are necessary for flight:

a) In static flight, the wind must not be flowing horizontally, but must have a vertical component.

b) In dynamic flight, the air flow must not be uniform but must vary in direction and velocity.

These two basic conditions make it clear that flight, both static and dynamic, is a question of atmospheric currents. Even today, however, gliders are still restricted to static flight which is based on the fact that the vertical speed of the glider is equal to or lower than the upward vertical component of air, thus permitting flight at constant or increasing altitude.

2. Concepts of Dynamic Flight.

Dynamic flight derives energy from variations in horizontal air flow. This is explained as follows: The lift of an airplane is proportional to the square of the speed of the air. Thus if wind speed increases, then relative speed increases, since the airplane, because of its inertia, does not adapt itself immediately to the new state of equilibrium, so that the additional kinetic energy makes it possible to gain altitude. If the airplane flies with the wind, relative movements are produced just as when the wind velocity declines.

These variations in flow may occur successively or they may be locally superimposed or opposed to each other. The energy required for flight may be drawn from variations in the velocity of horizontal flow, if great enough and several seconds in duration, so that velocity variations can be compensated for in the airplane by the appropriate pushing forward of the stick.

For simplicity's sake, we can restrict ourself to dynamic flight which utilizes local wind variations. In the lower air layers close to the ground the wind velocity increases noticeably with altitude. This variation might be used for dynamic flight by pulling back on the stick, in the case of a headwind, to convert the higher air speed into altitude and then making a 180° turn in a downward glide with tailwind, and so forth (Figure 1). The Albatros of the South Seas apparently flies in this manner.

Between 1923 and 1929 P. Idrac carried out detailed research on the flight of the Albatros and other sea birds. According to his findings, the albatros, petrel and other birds frequently fly continuously the whole day long, if the wind is strong enough, according to a definite pattern and can cover great distances without a single flap of their wings.

They start at the level of the waves, usually to the lee of a wave, and climb headwind to an altitude of 10 to 15 meters; then they turn into a position with tailwind or crosswind and glide down to begin the same maneuver again at sea level. The period of this maneuver varies with the species and flight path of the bird. The quicker the bird flies into a position with tailwind and the less altitude it gains, the shorter the duration of the maneuver. The albatros performs this maneuver in approximately 11 seconds, with the highest altitudes attained less than 20 meters. The albatros' minimum speed at sea level is 5 m/sec. The energy necessary for this dynamic flight is obtained from the gradient of the wind in the lower levels of the atmosphere. According to measurements, the wind velocity is roughly doubled at a 0.5 to 20-meter altitudes above the wave crests of a slightly rough sea. The wind velocity decreases considerably between the waves, so that the wind variation with altitude is increaded still more.

It also happens frequently that wind velocity is increased in one certain spot, so that the air masses of the higher faster-moving layers break through the lower ones. Thus, static and dynamic effects oppose each other and reduce the chances for dynamic flight.

At higher altidutes, great well-defined superimposed discontinuties are too rare to be of interest. However, it seems that opposing flows with distinctly different velocities may be utilized. These discontinuity layers may arise over terrain ridges, but, to be truthful, it is rather unlikely that the difference in velocities is great enough to create a noticeable effect in which the arising turbulence would allow the flying of a patterned spiral against the wind.

Dynamic flight is of no practical importance in human glider flight, since the use of the tactics employed by the albatros in the ground boundary layer would entail the performance of extreme and uninteresting aerobatics.

3. The Different Kinds of Static Flight

During its irregular progress, glider flight has opened three possibilities so far: They are:

Flight over slopes

Thermal Flights

Wave Flights

Flight over slopes utilizes the air current's upward vertical component created by an obstacle on the ground. The upwind over a slope is a local phenomenon and is limited to the flow created by the obstacle. The upwind over a slope draws its energy from the horizontal gradient of atmospheric pressure.

Thermal flight uses the upward thrust provided by air masses whose temperature is higher than that of the surrounding cooler air masses. The energy of a thermal updraft is merely transformed solar energy. Thermal flight is thus dependent on diurnal and annual variations in insolation and is also affected by geographical latitude.

Wave flight utilizes the ascending slope of an air flow in the form of a stationary wave which develops downstream from an obstacle on the ground. The energy required for wave formation is taken from the disturbance of the flow above and behind the obstacle. At the present, wave flight is still contingent upon topography and is limited to high rises on the ground.

- END OF CHAPTER 1 -

Chapter 2

Soaring Above Slopes

A. Air Flow Above Slopes

1. - Description of the Phenomenon.

Static flight was first effected by soaring. Since static flight requires an air flow with an upward component, the simplest method apparently is to look for the upwind required for a glider in the deviation of the air flow on slopes. Knowledge of air flow over an obstacle is, therefore, essential for soaring.

When air flows over an obstacle, its velocity increases with decrease in cross section of flow (Figure 3a). The air flow does not usually have the potential laminar and symmetrical form shown in Figure 3a. Because of increase in velocity, the static pressure is at a minimum directly above the crest; hence, before reaching the crest, the air pressure is diminished. The updraft is accelerated and clings to the slope. On the other hand, behind the slope the static pressure increases in the direction of the flow; the flow is consequently slowed down and may even be reversed. The counter-current forms an eddy behind the crest, which grows and separates from the obstacle (Figure 3b). New eddies form and separate, so that, behind the obstacle a row of eddies is formed which separates the zone of dead air from the undisturbed upper air flow. (Figure 3c)

This example indicates to us the basic currents to which attention must be paid in glider flight. Windward of the obstacle, that is, upstream, there is an updraft zone on the slope which is advantageous for glider flight. Directly above the crest, the curvature of the flow lines decreases with altitude and hence the upward component also decreases, so that the altitude attainable is limited by a glider's sinking speed. Leeward of the obstacle is a large downwind area full of eddies and with a zone of dead air in the lower layers.

The multiplicity of contours naturally complicates the types of air flow. This is shown, for example, by the flow lines of the air currents on Heligoland obtained hydrodynamically (Figure 4). In this sketch we can see the stationary eddies to the windward and leeward breaking away from the plateau of the island. In view of the steepness of the slope, the air flow to windward is slowed down to such an extent that on this side of the obstacle the gradient of the pressure is retrogressive. This causes the eddy on the windward side.

These types of air flow illustrates a two-dimensional air flow over the top of an obstacle. But to judge the strength of the updraft it is extremely important to know to what degree the air flow is due to passing over or around the obstacle; that is, whether there is a horizontal as well as a vertical deviation. We can expect an appreciably two-dimensional air flow; that is, passage over the obstacle only in case the slopes are highly developed and normal to windward. If there is only a slight slope it is necessary to pass around as well as over the obstacle, and if the gradient is great compared to length, passage around the obstacle is predominant.

2. Theoretical Determination of the Upwind Field

J. Ackeret (1) has pointed out a simple method, using the source method, of determining horizontal and vertical velocitites in the upwind field of an ideal obstacle. The terrain is replaced by a source on which parallel air flow has been superimposed. The resultant air flow supplies curves which represent the profile of the obstacle and the flow lines produced by the obstacle. This demonstrates that the geometric loci of equal horizontal and vertical velocity in the field are circles all of which pass through the source (Figure 5). The flow line which passes through the col-point P represents the profile of the obstacle. The velocity components can be easily read off at any point of the field.

The maximum updraft is directly above the source. The altitude which can be attained in a glider is:

$$H = \frac{Vh}{\pi v_z}$$

where h is the height of the obstacle;

V is the velocity of the parallel air flow;

v_z is the vertical velocity of the glider.

For v_z = 0.7 m/sec, V = 10 m/sec and h = 50 m, we obtain H = 227 m. This method by Ackeret assumes a potential, non-turbulent, frictionless flow, which of course is not found in nature. Thus it is not surprising that practical measurements do not coincide with those of the theoretical field.

3. Various Experimental Investigations

a) P. Idrac studied the distribution of upwinds over a range of hills southwest of Biskra. He found the values in Figure 6 for a horizontal velocity of 10 m/sec. The vertical velocity reached a maximum of 4 m/sec in front of and above the summit. It follows that the profile of the slope has a fundamental effect on the distribution of vertical velocities. Behind the summit, according to Idrac's description, is a great turbulent zone at the limit of which the undisturbed flow continues to rise. An upwind has been shown to exist close behind the crest.

b) Idrac also determined the upwind field of the gliding terrain of Vauville, at the tip of the Cotentin peninsula. According to his measurements, the distribution of rising velocities coincides well with the computations of Ackeret. Just as in Ackeret's representation the strongest upwinds over the Vauville cliffs are found directly over the slope. At an altitude five times the height of the cliff, measured from the foot of the cliff, Idrac still found an upwind of 0.5 m/sec for a wind of 10 m/sec horizontal speed, which fact agrees with theory.

The Vauville area has good upwinds, undoubtedly because it is under the influence of an undisturbed sea wind and is closed in by two Capes, Cape de la Hague and Cape Flamanville, which c annel the wind so that it cannot expand latterally and thus is forced to climb the cliffs.

c) Over the gliding terrain of the Polish city of Bezmiechowa, Kochanski has made very successful and instructive measurements of the upwinds there. These measurements differ from the above-mentioned ones because of a "cushion" of dead air at the windward side, so that the ascending field becomes noticeable only from a certain altitude on. His measurements of the upwind are particularly interesting. Figure 8 shows the distribution of vertical air currents in this zone for a wind of 8 to 11 m/sec. It can be seen that the obstacle throws a "shadow" and that the zone favorable for flider flight extends to far behind the crest. If strong winds prevail (8 to 10 m/sec), a large flat eddy can be observed; the downward velocity between points 1 and 2 is 4 m/sec.

d) Dunes with a regular profile and uniform ground are particularly suitable for studies of upwinds and flows in the presence of obstacles. The following results were obtained by the kinematographic recording of artificial smoke. The wind velocity is 15 m/sec. Its variation is nicely shown by the deformation of the smoke clouds (Figure 10). The turbulent layer separates two distinct air flows from each other (Figure 11). Above the turbulence is a uniform almost laminar flow, while the turbulent flow below is slowed down. In this difference between the two flows, one can recognize the friction between the ground and lower air layers.

This lower air flow, which is disturbed and slowed down, constitutes the boundary layer of the ground. According to actual measurements, it can be stated that the vertical velocities near the ground are greater than those determined theoretically. Without going into the other reasons, these divergences between theory and practice can also be explained by the terrain forms. According to calculations, a good glider should reach an altitude of 250 m above the dune; but according to practical measurements, it should rise a good deal higher than that. In a wind of 10 m/sec, trigonometric measurements were performed on a glider with a sinking speed of 0.6 to 0.7 m/sec. Its maximum altitude was measured at 250 m, in good agreement with theory.

B. LEEWARD FLOW

As far as leeward flow is concerned, theory, experience and practical observations are in agreement. Because of reduction in velocity, the flow is reversed in the turbulent leeward field. These leeward turbulences occur not only in wind tunnels, but also in the atmosphere. They may be stationary and they may be carried along by the flow. Figure 12 shows a stationary eddy as recorded by the flight path of a statically balanced balloon behind the raised dunes.

The flow to the lee of mountain chains is of importance only in wave flight. In true soaring, the leeward zone will pull the glider down and should therefore be avoided. The upwind zones discovered in glider flight in the undulating flow above mountain winds, which have been so successfully utilized, will be described in Chapter IV dealing with wave flight.

We must still deal with a vertical deviation of flow lines which, besides the unevenness of terrain, produces an upwind field analogous to that due to terrain obstacles, but much weaker. This phenomenon is called friction upwind.

C. FRICTION UPWIND

1. General

All fluids, including air, possess viscosity, that is, fluid friction. If an undisturbed fluid passes over a surface, the droplets next to the surface will adhere to it because of friction, and then because of their viscosity they will also slow down the neighboring droplets. Under the influence of that friction, a boundary layer is formed in whose interior the velocity passes rapidly from zero to the velocity of the nondisturbed flow. Thus the boundary layer is larger in dimension, the greater the viscosity and velocity of air flow and the greater the distance along the surface.

2. The Ground Boundary Layer

The viscous phenomenon also occurs in the flow of air over the ground. To be sure, the relationships here are much more complicated, since air friction in the atmosphere is determined not only by internal molecular friction but also by the continous turbulent mixing of the air particles.

The ground boundary layer is quite large. While the aerodynamic boundary layer, which is of fundamental importance, is only about one mm thick, the ground boundary layer reaches a thickness of about 100 m.

Naturally, the differences in the ground's roughness slow down the air masses in varying manners and thus cause variations in the extent of the boundary layer.

Air passing from a zone with little friction to one with much greater friction slows down and thus causes a congestion of air which the adjacent air flow must pass over; it is due to this phenomenon that we can observe "friction upwinds" at the borderline of terrain zones of different roughness.

These friction upwinds are usually found along the coast, at the borderline between water and land, and also over land at the place of transition from level ground to a built-up urban region. The upwind zone is marked by an increased boundary zone upstream from the point of change in roughness of the terrain. We can see this boundary layer even in Figure 11, where the top of the slowed-down boundary layer is marked by the layer of characteristic strong winds.

In the boundary layer, the horizontal velocity of the wind increases with altitude from zero velocity at the ground to the velocity of the general flow. Karman has worked out a simple procedure for determining the vertical velocity of a fraction upwind. The following relation can be established between vertical velocity V_z, horizontal velocity V_x, altitude H, and distance x from the start of the rough zone on the ground:

$$V_z = \frac{H V_x}{2x}$$

If we have, for instance,

$x = 1000$ m, $V_x = 20$ m/sec, $H = 200$ m, then it follows that

$V_z = 2$ m/sec.

Thus in order to make use of friction upwinds in glider flight, the winds must be quite strong. Furthermore, the upwind zone that can be utilized is limited; e.g. to a narrow strip along shore. Therefore the friction upwinds do not seem to be of immediate interest for glider flight; nevertheless this type of upwind should be known for possible future utilization.

D. THE GENERAL INTEREST IN SOARING

Flight in the upwind of obstacles has brought about a very careful study of the effect of complex ground configurations upon air flow and has slowly given rise to the development of meteorological aerodynamics. At present, soaring does not yet hold a preeminent position, but it would be false not to recognize its importance.

For glider training, upwinds are always a great help.

On the other hand, during long-distance flights, even in favorable weather, the pilot will be lucky to find even one upwind while waiting for the development of a new thermal gust.

Likewise, in discussions of thermal air currents, we will often come back to the preceding statements on the phenomena of air flow over terrain obstacles, since air flow over ground will obviously influence the generation of thermal upwinds. Furthermore, it is important for all pilots, whether flying powered or motorless airplanes, to study the lower layers of the atmosphere which are influenced by ground contours; that is, they should study the atmospheric "surf", in analogy to the sea surf.

At a certain altitude, these "storms" of the aerial "ocean" will not affect an airplane; but close to the ground, in the "surf" zone, they may have serious consequences. The study of flow forms in this important zone allows the pilot frequently to make rough flights more pleasant and often to avoid danger.

The preceding statements dealing with the effect of mountainous obstacles on air currents will now be illustrated by an example of their practical utilization, namely long-distance pure soaring flight. Figure 15 shows very nicely that such distance flights are flights from one crest to the next, in the course of which the pilot gains altitude while flying over normal slopes with a wind which blows across and not around them. The upwind is greatest when the gradient of the slope is greatest; i.e. when contour lines on the map are closest together.

Such a long-distance flight is a constant fight against the ground. In fact, continued mountain chains are rare and in flying from crest to crest the upwind zones are frequently separated by large downwind zones. It is a matter of skill and knowledge for the pilot to know exactly his relative altitude and the extent of the downwind zone in order to be able to reach the next slope.

For soaring, the pilot needs two qualities: He must have perseverance to wait in an upwind zone until the sailplane has reached the necessary altitude to fly to the next slope and he must also be capable of making quick decisions, so that he will leave the upwind zone where he is safe at the moment when he has reached the necessary altitude.

Furthermore, upwinds are also subject to the effects of night temperature and become considerably weaker over a continental mountain chain, even though the wind might increase. This has a very simple cause: During the night the valleys around the mountains are filled with immobile cold air which in a sense flattens out the contours, so that the slopes are completely submerged in a cold air mass or emerge only a little, thus considerably weakening the force of the upwind.

For the best soaring conditions, the most suitable terrain is the seacoast. During the night the sea is comparatively warm and the horizontal velocity of the wind is greatest, so that most favorable conditions for soaring are established. It is thus best to seek the regular dunes parallel to the seacoast for setting endurance records in gliders.

- END OF CHAPTER 2 -

CHAPTER 3

THERMAL GLIDER FLIGHT

1. General Principles.

We shall now describe the first thermal flight, 30 April 1928, not so much to recall this historic event as to point out all the characteristics exemplified in this flight.

This first thermal flight was of the greatest importance, because it confirmed the newly discovered possibilities of glider flight and because it was the true starting point of motorless flight.

Two important observations on cumulus clouds were the stimulus for this first tentative thermal flight: namely the observations that clouds form and dissociate at intervals of 25 to 30 minutes and that this phenomenon is repeated always at the same point with respect to the sun's position. Hence it was concluded that cumulus clouds and, even more so, upwind fields have a stationary development during a certain period. On the basis of these promising findings, the pilot Johannes Nehring was ordered to fly a light 35-HP airplane to the base of a cloud and to describe wide circles with a slight bank and with the engine shut off. Considering the era during which the experiment was performed, the results were astonishing: The airplane did not lose altitude and succeeded in remaining aloft underneath the cloud. Thus the first human thermal flight had been carried out. The barogram of the flight (Figure 17) confirmed the visual observations. It is shown that the airplane carried out a short glide after a normal ascent and with the engine off, and that it could maintain itself-between points 7 and 10 on the barogram - in the thermal updraft of the cloud-without losing altitude for 9 minutes. Thus, the airplane's sinking speed being known, the vertical speed of the air underneath the cloud could be determined. The upward speed turned out to be 4 to 5 m/sec. These speeds are rarely attained in upwind fields above slopes. Thus was demonstrated immediately the great superiority of thermal soaring over soaring above slopes.

2. Equilibrium conditions of the Atmosphere.

The vertical speed and the development of thermal motion at high altitudes depend on the momentary state of equilibrium of the atmosphere.

In order to have stable air, it is not enough that the upper layers be less dense than the lower layers, for the relation between density variation of an air mass in vertical movement and density variation of the surrounding still air will also have an effect.

Three states of equilibrium of air can be distinguished:

a) Stable equilibrium: Air mass lifted vertically must return to its original position, since its density variation with altitude is less than that of the surrounding air. When rising, the air mass becomes heavier than the surrounding air and is therefore held back; in descending it becomes lighter than the surrounding air and generates an ascending force.

b) Neutral equilibrium: An air mass lifted vertically can come to rest at any altitude, because its density varies at the same rate as that of the surrounding air.

c) Unstable equilibrium: An air mass lifted vertically continues in its original motion and is accelerated, because its density gradient at various altitudes is greater than that of the surrounding air. The air mass becomes lighter than the surrounding air masses during its ascent, and heavier during its descent.

Depending on the direction of the vertical motion, any air mass is influenced by increase or decrease of pressure and will expand or contract adiabatically, as the case may be, without taking up or giving off heat.

But this air mass, because of the effect of the loss or gain of energy during compression, will cool off or heat up. Heating and cooling occur according to the adiabatic law and correspond to a temperature variation of $\Delta T = 1^{\circ}C$ per 100-meter change of altitude. Thus, in neutral equilibrium, if the air molecule set in vertical motion is subjected to the same density variation as the surrounding air, it will also be subjected to the adiabatic laws and show a temperature gradient of $1^{\circ}C$ per 100 m. Therefore, by means of the vertical temperature gradient, the equilibrium conditions can be classified according to the following manner, as shown in the numerical table:

Temperature in °C

Altitude in m	Still air	Vertically moving air	Still air	Vertically moving air	Still air	Vertically moving air
1500	2	5	5	5	12.5	5
1000	8	10	10	10	15	10
500	14	15	15	15	17.5	15
0	20	20	20	20	20	20
	Unstable equilibrium		Indifferent equilibrium		Stable equilibrium	
	$\frac{\Delta T}{\Delta H = 100\,m} > 1^{\circ}C$		$\frac{\Delta T}{\Delta H = 100\,m} = 1^{\circ}C$		$\frac{\Delta T}{\Delta H = 100\,m} < 1^{\circ}C$	

In an unstable equilibrium, the temperature decrement for still air has been assumed 1.2°C/100 m. An air molecule lifted by the force of a slight gust or by a **terrain obstacle** has a temperature gradient of $1^{\circ}/100$ m. Thus, at 500 m altitude it has acquired an excess temperature of 1°; at 1000 m, 2°; and at 1500 m, 3°. Therefore the ascending speed of the particle increases with altitude. This explains the great possibilities inherent in thermal flight.

In the example of neutral equilibrium we find that the temperatures of still and moving air are the same at all altitudes. There is thermal equilibrium at any altitude, and vertical air movements are dissipated through friction and can stop at any arbitrary altitude.

In the case of a stable atmosphere, the temperature gradient is 0.5° per 100 m. A molecule in motion will cool off by 1°C per 100 m; i.e. with respect to the surrounding still air, it becomes colder when rising and warmer when falling. At an altitude of 500 m, the negative difference in temperature has already reached 2.5°C; the particle will thus return to its initial position at ground level due to its negative acceleration. In stable equilibrium, the vertical movements which have been started are thus slowed down and the free vertical movements of air will produce distinct conditions only within a limited range. The chances of thermal flight under stable equilibrium are therefore rather slim.

Temperature-altitude diagrams show very clearly the state of equilibrium of the atmosphere and the vertical movements of isolated air masses. These graphs of the atmospheric state picture well the thermal conditions and the corresponding prospects of thermal soaring.

In a coordinate system, the temperatures are shown on the abscissa and the altitudes on the ordinate. If we draw the graph at a scale which will show 1°C of temperature and 100 m of altitude by the same unit of length, adiabatic temperature distribution will be represented simply as a stright line with a 45° slope (Figure 18). For this gradient, still air and individually moving molecules have the same temperature at all altitudes, i.e. the atmosphere is in a neutral state of equilibrium. In Figure 19, a stable equilibrium with a ground temperature of 20°C is represented. Temperature variation with altitude of the still air is represented by straight line T. We now assume that a small isolated air mass is lifted from the ground. In the course of its ascent, it will undergo adiabatic cooling of 1°C per 100 m. Temperature variation with altitude will be represented by the adiabatic line T'. We find now that the ascending air mass is colder and heavier than the surrounding air and that a return force acts upon it. This return force increases with altitude. Thus the vertical motion of the air mass is rapidly stopped and, because of the descending force, it will again sink down to the ground.

At the moment when the air has the same temperature, be it moving or still, there is thermal equilibrium and, until vertical movement is resumed, the atmosphere is in stable equilibrium. In this case, the adiabatic line will be left of the curve of still air.

In an unstable condition, the phenomenon is reversed (Figure 20). Let $\Delta t/\Delta H(= 100\text{ m})$ be 1.2°C, a vertical gradient of temperature. If an isolated air mass if lifted slightly from its rest position, its temperature will vary adiabatically with altitude, becoming warmer and therefore lighter than the surrounding air, and will therefore be acted upon by an ascending force which increases with altitude. The atmosphere is therefore in unstable equilibrium and the adiabatic line will be right of the curve of still air.

In neutral equilibrium, the temperatures T and T' will coincide and thus form one line, the adiabatic line $\Delta t/\Delta H(= 100\text{ m}) = 1°C$.

These simple graphic representations can always be used to indicate the various possibilities offered by vertical thermal currents.

3. The thermal energy sources of the atmosphere.

These considerations of the equilibrium conditions of the atmosphere have given us an insight into the physical bases of the vertical motion of air currents. In order to solve the problem of thermal flight, we must now find out which are the sources of thermal energy which the atmosphere has placed at our disposal.

We shall proceed from the basic idea that our atmosphere is a thermal machine with a hot and a cold source. It exchanges horizontally quantities of heat, between the tropical and polar regions, in the form of hot air masses from the tropics advancing toward the North and cold polar air masses advancing toward the South.

While this horizontal heat exchange is of great importance for the temperature prevailing on the earth, it is of only limited interest for glider flight at the present time.

The flow over mountain obstacles which gives rise to updrafts over slopes is the result of this horizontal heat exchange and of the pressure differences caused by it.

However, the thermal updrafts are caused by another kind of function in this atmospheric thermal machine.

If we recall that freezing temperatures are encountered at 4000 m altitude in mid-summer, it is easy to realize that there is a local exchange of energy along a vertical line between the hot source at ground level and the cold source at high altitude.

The kinetic energy of the vertical motion of air which is utilized in soaring represents simply the transformation of the potential energy of the air column above a given point. This potential energy is the greater, the greater the heating below that point and the greater the cooling above it.

These two phenomena, the heating of air at ground level and the cooling of air at high altitude, impart to the air unstable equilibrium and cause the potential energy of the still air to be transformed into the kinetic energy of the vertically moving air.

How do the heating and cooling of the atmosphere and the creating of vertical movement of air masses take place? This is the fundamental problem of thermal flight.

The unstable temperature distribution which makes the creation of updrafts possible can be the result of various phenomena:

1. The heating of the ground and the adjacent air layer by direct insolation. If we eliminate cloud formations, this phenomenon leads to the formation of pure thermal currents, by which term we mean thermal updrafts occurring in cloudless weather.

2. The arrival of water vapor produced by evaporation on the ground or by advection, i.e. the arrival of an oceanic air current. The heat of condensation liberated by the formation of clouds frees a great amount of potential energy which makes the atmosphere unstable and creates the clouds updrafts well-known to glider pilots.

3. Cooling at high altitude, either by radiation of higher and more humid air layers or by advection of colder air at high altitude. These phenomena are independent of the time of day and can produce thermal updrafts also during the night. Since in this case the updrafts are produced at a certain altitude, regardless of ground conditions, they are called high-altitude thermal currents. In contrast to thermal currents caused by heating, these high-altitude thermal currents are caused by cooling and are of no great importance.

4. Excess temperature of a water surface as compared to the surrounding air. This creates the marine thermal currents. While the physical principles of marine thermal currents are fundamentally the same as those of thermal currents over land, it seems advisable to treat thermal currents over water separately, since the oceans take a special position in regard to glider flight. This thermal current is produced when the water temperature is much higher than the temperature of the surrounding air. In our geographical latitudes, this occurs generally during winter. These conditions are also fulfilled in the event of an outbreak of cold polar air.

In the vast expanses of the tropical seas, especially in the trade wind zones, the water surface is warmer than the air throughout the year, giving rise to marine thermal currents.

4. Thermal flight without clouds. (Solar thermal current)

Thermal flight in a clear cloudless sky is doubtlessly a difficult problem in glider flight. Since the updraft zones are not marked by clouds, the pilot must locate them entirely on his own, aided only by his hunches and by the indications of the variometer. Especially in this case, the personal knowledge of the principles of thermal updrafts is of great help.

To treat the problem of the dry unstable state of equilibrium of the atmosphere, we must first ask ourselves, keeping in mind the conditions of equilibrium discussed above, whether there are any thermal updrafts in a dry stable equilibrium $\Delta t / \Delta H (= 100) = 1°$. It can be stated in answer to this question that there are possibilities of gliding in an stable atmosphere. However, the updrafts are isolated, occurring only above favorable terrain features, and do not rise very high. Let us imagine a clear sunny morning on a dry lawn-covered airfield surrounded by woods and fields. The air temperature is 20°C. The air heats irregularly. Over the fields, let us assume, there is still a temperature of 20°C. The dry sunny airfield will heat more quickly; there the air temperature will be 23°C. In the woods, the air is still cool and has a temperature of only 18°C. Since there are different temperatures on the same level, there is no equilibrium. In comparison with the surrounding air at 20°, the air over the airfield has an excessive temperature. It will leave its stationary position and rise. Figure 21 shows the vertical extent of this local updraft. In still air, the temperature gradient between ground and 1000-m altitude is 0.8°C per 100 m; between 1000 and 2000 m altitude the gradient is 0.6°C per 100 m. The rising air, with a temperature of 23°C at the ground, will be adiabatically colled by 1°C per 100 m.

A comparison of the temperature curves of the rising air and the surrounding still air shows at once that the air over the airfield, overheated by 3°, can rise to an altitude of 1200 m. Up to that altitude it possesses an ascending force and its motion is accelerated. Since it has its greatest vertical speed at an altitude of 1200 m, it begins to go beyond its altitude of thermal equilibrium. But the return force slows it down rapidly and it returns to its altitude of equilibrium. Further examples will show that the highest altitudes attained by gliders, when their sinking speed is taken into account, correspond very closely to the altitude of equilibrium of rising air masses. Therefore it is unnecessary in practice to pay much attention to air movement above the altitude of equilibrium. Local increases in temperature, caused by the uneven heating of the ground, will create isolated thermal updrafts of limited altitude in a stable atmosphere. These conditions characterize the early morning hours of clear summer days.

Let us now deal with the updraft conditions around noon or the early afternoon of a hot summer day. In this example, the formation of clouds is not necessary. The lower air layers are generally overheated by strong insolation; the atmosphere for several hundred meters above ground is thus unstable. In the temperature-altitude diagram of Figure 22, the still air up to 500-m altitude has the temperature gradient $\Delta t / \Delta H (=100) = 1.2°C$; from 500 to 2000 m the equilibrium is stable with a temperature gradient of 0.8° per 100 m. An isolated air mass which is rising adiabatically is then carried to an altitude of 1400 m. If the isolated air mass has a temperature excess of 2° at ground level, it will rise to 2000 m. Since dry unstable temperature gradients are found only up to a few hundred meters altitude and do not greatly exceed 1°C, these thermal updrafts without clouds generally do not rise very high. Although the instability of the lower layers gives rise to the condition which causes free updrafts, it is of importance only near earth's surface. There develops an interchange of updrafts and downdrafts with no regular distribution, requiring great experience and skill in flying. The ascending air forms a limited laminar flow. The diameter of the updraft zone is of the order of several dekameters. For reasons of continuity, a downward movement must correspond to each ascending current. This downdraft generally has different characteristics from the updraft, being not a limited movement, but an irregular downward movement of air masses extending over wide areas and progressively slackening during the descent.

Even in unstable equilibrium, an exterior stimulus is required to initiate the vertical movement, because up to a gradient of 3.4°/100 m the decrease in pressure will exceed the decrease in temperature; thus the air will remain statically in state of equilibrium. However, in this unstable atmosphere, one isolated particle will receive a chance impulse through turbulence or from a small terrain obstacle, and this negligible initial movement will trigger or release a huge vertical movement of air.

Figure 23 shows a rapid recording of the temperature in air layers next to the ground, between the altitudes of 4 and 80 m. It is seen that an extremely high gradient may occur between 4 and 15 m, of the order of 4° per 10 m. In the free atmosphere, such high values are never attained nor any nearly that high.

This highly overheated lower layer is called the "thermal boundary layer" of the atmosphere. It is the zone which is the origin of the updrafts observable on the ground. Every automobile driver knows this layer in the form of shimmering layers of hot air above the highway on hot midsummer days. This boundary layer which can be seen in Figure 23 extends to an altitude of about 15 m. It is also evident from this illustration that updrafts try to burst out through the upper limit of this boundary layer. Highly overheated air masses at the ground detach themselves from the ground at regular intervals and begin to rise. Their place is taken by colder air masses which descend.

The variations between 22 and 26°C in the temperature of the boundary layer show that it is cooled from above and then heats up again until a new updraft of hot air is released.

Figure 24 shows the formation of the cold boundary layer, which takes place shortly before sunset when the ground is no longer insolated. In this illustration, a feeble updraft can be recognized at 1510 hours; afterwards there are no more thermal currents and the temperature gradient is modified.

Wind recordings also show very clearly by, variations in force and direction, this periodic release of the boundary layer.. Figure 25 shows a recording with practically no wind until 1130 hours. At that time the cycle of thermal emissions begins. It can be seen from the more or less regular wind variations that the rising air of the overheated boundary layer is replaced by a discontinuous influx of fresh air in a horizontal direction. On the basis of absolute maximum wind velocity, and average period of 25 minutes is obtained for the release of updrafts. The variations in wind direction correspond to the variations in wind velocity. The fresh air which flows in horizontally from the surrounding area arrives once from the North and once from the South. In the example of Figure 26, the recording of wind direction shows the periodic emissions more clearly than does wind velocity. The mean wind velocity, in this measurement, is rather high and occasionally reaches 10 m/sec. The emissions are not as frequent, but they are marked more sharply. During each emission the wind shifts from East to Northeast or to North. The period of emissions here is approximately 38 minutes.

The flow of thermally rising air above the boundary layer is excellently demonstrated in desert areas where the boundary layer sparkles and reflects like the surface of a large lake. Suddenly there rise from this layer narrow sandspouts, a striking spectacle which enlivens the monotony of desert landscapes.

During the dog days, these little whirlwinds occasionally develop also in our climate above extended overheated areas with cooler zones nearby. One of these miniature whirlwinds was detected by pure chance during measurements performed with a statically balanced meteorological balloon, as shown in Figure 25. It demonstrates clearly the discontinuous variations in wind velocity and direction, characteristic for thermal currents. During the 14th minute of the recording, the balloon was seized by an ascending flow and carried from 500 to 1100 m altitudes; and after several minutes of calm, a rapid downward flow arose, with uncommonly high vertical velocities. These movements of the balloon can be considered as conforming to Letzmann's pattern of small whirlwinds, with the air on the outside rising in a sinusoidal spiral and a downward compensating flow through the axis of the vortex. There is also a report available on the meeting of a glider with such a whirlwind. The pilot observed a dextrorotatory whirlwind, half a meter in diameter, 100 m below the base of a cloud, which he could not avoid. His left wing was caught in this whirlwind and raised, so that the airplane was suddenly stood on its wing tip. According to the pilot, only the sturdiness of the airplane prevented a crash.

The phenomenon of flow above the boundary layer can also be shown experimentally by setting up smoke pots in a big circle in a strongly overheated area. At first the smoke emitted will flow arbitrarily in all directions, but suddenly the smoke from all pots will converge in one definite point within the circle and rise in a continuous column to a certain altitude. One of these experiments showed that the smoke flowed toward a cloud in the process of formation at that time at an altitude of 1500 m, whereby all factors in the development of thermal updrafts were demonstrated.

It has been emphasized above that an unstable air layer requires some initiative impulse before it can rise freely. Isolated air masses must undergo a slight vertical disturbance by an external impulse in order to continue rising freely under the influence of unstable stratification.

This release of unstable stratification can be initiated by:

a) terrain obstacles (towns, forests, hills, mountains)

b) turbulence in the surrounding air

The release of thermal updrafts can be observed over any terrain obstacle, mountains being the most effective one.

The air flows against the mountains and is forced to rise along the slopes. In an unstable atmosphere, this upwind changes into a free thermal updraft. Under these circumstances, the upwind grows and reaches vertical speeds quite different from those of the simplified discussed flow phenomena corresponding to the flow of incompressible fluids.

Figure 28 shows the barogram recorded in a glider. It clearly shows a pure upwind above a slope and the transition from this slope upwind to a thermal updraft.

The release of unstable air masses is a phenomenon which has been studied by numerous measurements carried out by gliders and statically balanced meteorological balloons. The emission phenomena over mountain crests or slopes are the most interesting. Because of the release of air in an unstable thermal state over the crest, the upwind zones are not strongest on the windward side of the slope; but, rather, the upward flow attains its greatest force on the leeward side.

The flight paths of balloons, shown in Figures 29 and 30, were recorded on the same day and over the same terrain but at different moments. The first measurement (Figure 29) was performed at 1107 hours, the second one (Figure 30) at 1135 hours. They show the same picture. The location of the upwind zone in both recordings is the same. A third measurement, performed at 1505 hours on the same day, shows a strong upwind zone to the leeward side. This series of measurements shows that more or less fixed upwind zones develop to the leeward of a mountain massif, in the same manner as the stationary upwind zones which we will finally encounter in Chapter IV in our treatment of flow phenomena associated with fixed waves.

The following flight paths (Figure 31) show the phenomena of release of the thermal boundary layer over flat terrain, especially at the edge of a forest or, in general, whereever there is a change in roughness of the groundlsurface. The flight paths of the balloons show that the overheated ground air is blown slowly toward the forest edge, following the wind direction and that it is lifted above the ground there to generate a thermal updraft above the forest.

Figure 32 is also very instructive. A balloon was released from an airplane at 500 m altitude. After its release, it began to drop and then remained for nearly ten minutes without any vertical motion. Then suddenly, 21 minutes after its release, the balloon rose rapidly at a speed of 2.5 m/sec from 300 to 1100 m. This ascent also took place above a forest edge. Obviously the following phenomenon has taken place: Upwind from the forest is an overheated boundary layer. A small gust pushes the boundary layer above the edge of the forest and releases the ascending flow which seizes the balloon of Figure 32 at an altitude of 300 m. A number of measurements performed under the same conditions confirm the existence of an updraft field over the forest.

Figure 33 shows such an updraft field downwind from a forest. Of course, the updraft is not stationary. However, it indicates that for a certain wind direction and strong insolation updraft fields are fairly frequent on the downwind side of forests. These phenomena of updrafts over mountains, hills, forest edges, and towns, are good means for locating thermal updrafts when there are no clouds to indicate their existence. If the wind's force at the ground is great, an undisturbed boundary layer will form only with great difficulty, since the turbulence of the flow will always remove new particles from the layer, thereby continously destroying it. Free updrafts will not develop, except as short choppy gusts.

Temperature inversions; atmospheric blocking layers.

In the above treatment of the subject it has been assumed that temperature decreases continuously with increasing altitude, which is the normal phenomenon of a cooling atmosphere. However, it is frequently found in the atmosphere that temperature decrease with increasing altitude is interrupted by a narrow layer whose interior has constant temperature or even increasing temperature with increasing altitude. The increase in temperature can be quite considerable, in the winter reaching as much as 10°C. These layers, in which the temperature increases with increasing altitude, are called temperature inversions. If the temperature remains constant with increasing altitude above a certain layer, we speak of isothermic conditions. These temperature inversion layers are very important. They are very stable and can thus slow down considerably the upward movement of air and prevent the air masses from rising higher. Therefore the inversion layers are also known as blocking layers. The inversions also represent the boundary layers which separate air masses of different properties (humidity, visibility, wind direction and force, etc.). Figure 34 shows an example of temperature inversion in which the temperature rises 3.5° between 1500 and 1800 m. In this example the air is dry and unstable between the ground and an altitude of 1000 m, with a temperature gradient of 1.1° per 100 m; nevertheless an isolated air particle rising from the ground having an excess temperature of 5° is prevented by the inversion from rising above 1700 m.

If there is little or no insolation, as between evening and early morning or in the winter, these temperature inversions are very common.

Inversions at the ground are formed in the lower layers, when the temperature of the soa becomes lower than that of the air. Figure 35 gives an example of this. Under such conditions, obviously, thermal updrafts originating at the ground cannot leave the ground. The adiabatic line, for an initial ground temperature of 15°, will always remain left of the curve of still air temperature. An excess temperature of 6°C is required before an isolated air mass can leave the ground.

The above example shows, above the inversion at the ground, a gradient of 1.1° between 500 and 1100 m, and thus a dry unstable state. There can thus be thermal updrafts in the zone between 500 and 1300 m (adiabatic line T"), as far as isolated air masses can be displaced from the equilibrium altitude. In this example, the updraft is initiated by a chain of mountains. In the case of temperature with unstable stratification, air lifted by this manner above 500m can freely rise.

We have here an example of mountain thermal updrafts in the evening which persist while daytime thermal currents and updrafts originating in the valley have already stopped long ago because of the evening coolness.

Figure 36 shows the destruction of the ground inversion by the progressive morning heating and the start of vertical thermal movements.

The ground inversion, about 50 m. deep, is attacked by the ground air masses which become overheated by growing insolation to progressively greater depths, until the inversion disappears altogether around 1100 hours. Then an adiabatic gradient is produced which changes to a superadiabatic gradient. From that point on, thermal updrafts can reach high altitudes. At the end of the afternoon, the radiation from the ground becomes stronger than insolation and thus a new inversion is formed. Upon this stabilization of the air, the thermal updrafts stop since they have been cut off from the zone generating them, and tend to disappear to higher altitudes. This phenomenon of the formation of inversions on the ground can be demonstrated by the rapid recording of temperatures between the ground and an altitude of 80 m. Figure 24 shows that there is still a thermal updraft traveling quickly to higher altitudes at 1440 hours. The following emissions of the boundary layer, between 1450 and 1510 hours, are already weakened considerably. Immediately after this last updraft, the inversion is formed at an altitude below 50 m, thus preventing the development of any further development ~~of any further~~ updrafts. At 1600 hours the inversion at the ground has been completely formed and the adjacent layers are in a thermal state of rest. It is astounding to see how upper layers react quickly on a thin, less cold air layer on the ground and how thermal vertical currents are stopped by the formation of an inversion at the ground. On some days, with full insolation, the stabilization of the lower air layers may be so strong as to stop the development of thermal updrafts altogether.

Examples of inversions:

a) Brazil: -

In the course of a glider expedition to Brazil, it was found on a number of days that the best chances for thermal flight at Rio de Janeiro, that is, in the immediate vicinity of the Atlantic Ocean, are between 900 and 1100 hours. Often, all chances of thermal flight are gone by noon, just at the time of highest sun. The reason for this surprising phenomenon is the transition from a land breeze to a sea breeze which occurs at noon. Night and morning, there is a slight wind from land which is caused by the strong heating of the continent. The land wind heats up quickly and strongly under the influence of the sun and offers excellent opportunities for gliding. The temperature curve, due to the land wind, is in an unstable equilibrium, in the dry state, up to 1100 m altitude; then its instability increases greatly up to the cloud ceiling. The conditions are altogether different, starting at noon, when the sea breeze blows in the Atlantic coastal regions.

A thin film of sea breeze slips underneath the mass of hot air and detaches it from the ground. There is now colder air underneath the higher, warmer air masses so that the atmosphere is completely stabilized. The diagram of the temperature of sea air (Figure 37) shows that the cooling extends approximately to 600 m and that its altitude is limited by a strong inversion. Cooling by sea breeze amounts to approximately 6°C. The adiabatic line is completely left of the curve describing the state of the air; thus stability is complete. Beside the impossibility of thermal gliding despite a temperature of 28°, the stability of the air is also shown by the complete absence of any gusts. With full insolation and with a 6 to 8 m/sec wind, it is possible to fly an airplane right next to the sheer rocky cliffs near Rio without noticing the slightest turbulence.

b) Libya

Analogous conditions can be found along the North African coast in Libya. There exists a great difference in the possibilities of glider flight between East wind and West wind. As seen from the coast, the East winds come directly from the sea, while the West winds have to traverse a stretch of land. Figure 38 gives the curve of mean temperature at Homs, 50 km

East of Tripoli, for East and West winds, as measured during research flights carried out between March 29 and May 5, 1939. These curves show that the East wind is definitely warmer than the West wind, especially above 1000 m altitude. However, the East winds' temperature curves shows an inversion at 1000 m, while the inversion layer for the West winds is as high as 1700 m. Thus the altitudes attainable by gliders are limited to 1000 m by the East wind, and to 1500 - 1700 m by the West wind. These altitudes are confirmed by actual flights performed, as shown in the maximum altitudes attained for a West wind on 2 April 1939 (Figure 39). At Garian, on a 700-m high plateau 120 km from Tripoli in the interior of the country, the sea breeze is no longer felt (Figure 40). The inversion altitude for the East wind is 2000 m, while there is none for the West wind. The soundings of 9 April 1939 (Figure 41) showsthe correspondence between altitudes reached by glider and the equilibrium altitude of rising air. The inversion being much higher, the possibilities of glider flight at Garian are much better than at the coast of the Gulf of Libya.

The barogram of 4 April 1939, (Figure 49) for a long-distance eastward flight from Homs across the Great Syrtis desert, shows very well the difference between the two meteorological regimes at the coast and in the interior. Near Misurata, the coast describes a right angle toward the South. South of Misurata, the West wind traverses a rather large stretch of land. Therefore, the barogram shows at this point a higher ceiling, while at Homs, under the influence of sea breezes, soundings of the temperature during the day indicate an inversion at an altitude of 900 m, namely the ceiling altitude for flights along the coast. South of Misurata, to the contrary, the continental influence is stronger and a glider can rise to 1700 m. All along the flight, altitude variations are slight.

This barogram is a good example of a flight underneath clouds or, perhaps, for the stationary waves along the coast. These aspects will be discussed below.

5. Cloud flights

Until now, for the sake of simplicity, we have considered equilibrium conditions of the atmosphere without taking into account the phenomena of condensation and cloud formations in the ascending flow. However, a rising air mass will generally cause the formation of clouds. These clouds are, on one hand, the most reliable indications of the existence of updrafts and, on the other hand, increase considerably the speed of the updraft.

When a cloud is formed in a rising air current, modifications in the thermodynamic state of the rising air become quite complicated. We have already noted that non-saturated rising air cools off by 1° per 100 m at any altitude. When the phenomenon of condensation intervenes here, these simple relationships change. Physics shows that the vaporization of water requires a considerable amount of heat. This heat of vaporization is contained in the water vapor in the form of latent energy and is liberated upon transition of the water from the gaseous to the liquid state (heat of condensation). Consequently, when adiabatic cooling of a non-saturated air mass causes condensation, the heat of condensation liberated thereby is absorbed by the rising air. Thus, the cooling of the air, which was 1°C per 100 m in dry air, becomes less because of condensation, so that saturated humid air which is rising has a temperature gradient below 1° per 100 m.

This gradient is not constant, but is a function of temperature and pressure, as is shown by the following table:

Altitude	Temperatures 30	20	10	0	-10	-20
0	0.37	0.44	0.54	0.62	0.75	0.85
1020	0.37	0.46	0.5 6	0.68	0.82	0.90
2000	0.38	0.49	0.60	0.75	087	
3000	0.40	0.53	0.65	0.82	0.89	
4000	0.42	0.57	0.73	0.88		

Thus, the equilibrium conditions previously given for dry air must be complemented by the relation for unstable humid equilibrium

This condition means that an air mass of gradient below 1° will remain stable as long as there is no condensation. Rising saturated humid air is thus no longer cooled adiabatically, but according to the adiabatics of humid air with a variable decrement of temperature.

Thus the adiabatic line for humid air is no longer rectilinear, but is a curvilinear function of altitude and pressure which approaches the adiabatic line at high altidudes. As an illustration of the importance of humid-unstable equilibrium, let us consider the example in Figure 21.

In order to rise freely in a stable atmosphere, an air mass requires excess temperature in relation to the surrounding air. For a temperature excess of 3°, this air mass will find its altitude of equilibrium at 1100 m if no condensation takes place. Now let us suppose that condensation takes place at 900 m, under the same temperature conditions (Figure 43). The air particle, which up to this altitude has followd the adiabatic line T', along its further rise will now cool off according to the adiabatic line for humid air T" and will reach its equilibrium altitude at 2100 m, the air having been lifted an additional 1000 m by the energy liberated during condensation of the water vapor. At the same time, during condensation, the particle will be accelerated. The upward velocity of the air increases considerably upon arrival in the cloud. Thus, cloud flight is a considerable improvement over pure thermal flight; therefore it is not surprising that the first thermal flight was in the updraft underneath a cloud.

Figure 44 pictures the vertical velocities in an unstable atmosphere. They are computed mathematically from the differences in temperatures T' and T" and the temperature T of the surrounding air. In the example given here, the atmosphere is stable up to an altitude of 1350 m and becomes progressively more unstable and humid above that point.

It was not the excess temperature, but a 900-m high mountain, which caused the rising of the flow which attained an altitude of 1300 m at the humid-unstable layer. Inside clouds between 1300 and 2400 m, the vertical speed goes from 1 m/sec to 6 m/sec. This is a real example and corresponds to the first glider flight in a cumulus cloud. The glider was flying underneath a cloud and was sucked up to rise 900 m in 3 minutes. The pilot, who had no special equipment and had never done any blind flying, lost control and did not recover until he had reached 1600 m.. Allowing for a sinking speed of 0.7 m.sec for the glider, this means that there was a vertical speed of 6 to 7 m/sec in the cloud.

This example teaches which terrain offers good opportunities for gliding. A combination of soaring over slopes and thermal flight is expecially interesting for training terrain. This combination is given in flat terrain which is bordered at the downwind end (with respect to prevailing winds) by a chain of hills, as is the case where a broad valley has one side exposed to the prevailing wind. In our climatic zone, elongated terrain running SW to NE is favorable, since slopes running aling this direction are exposed to prevailing winds and are insolated for long periods.

We shall see below that terrain of this type facilitates gliding during the winter when there are no thermal updrafts, by the utilization of lifting waves.

We cite as an example the region of Mantes which fulfills the above conditions very well.

Let us discuss three different cloud flights performed on the same day and in the same spot. The temperature diagrams show an almost indifferent equilibrium below the clouds and, starting at 1400 m, a humid-unstable condition which increases with increasing altitude. The highest point of the cloud is at 3000 m. The flights on that day are quite typical illustrations of vertical-speed distribution.in an average-sized cumulus cloud. Although the flights were made at different times, between noon and late afternoon, various general statements can be made on the vertical air currents. Glider I (Figure 45) climed almost uninterruptedly from the airport at 900 m altitude to an altitude of 3000 m, at first in an upwind over a slope and later, starting at an altitude of 1500 m, in a cloud updraft. The glider flew through the cumulus cloud in a continuous climb. The maximum climbing speed of 5 m/sec was reached between 2200 and 2700 m. Glider II also flew through a cumulus which was not as large because of the late afternoon. In the barogram (Figure 46) of glider II it is particularly interesting to find that the airplane encountered violent vertical gusts at 1800 m. Within a few seconds the airplane dropped 140 m, only to climb 170 m immediately afterward. Two other gust of less violence followed. An examination of these gusts shows that they had a downward speed of 9 m/sec and an upward speed of 10 m/sec. The barogram (Figure 47) of glider III also shows violent fluctuations in vertical speed between 1200 and 1500 m. The vertical speed jumped within a few seconds from +4.1 m/sec to -0.8 m/sec, then again to +4.7 m/sec and then to +1.8 m/sec.

The following general deductions can be made on the basis of these cumulus flights:

1.) Very high skill and through knowledge of blind-flying instruments are required of the pilot so that he may recognize immediately the unnatural flight attitudes which his glider may assume.

2.) The glider must have good performance qualities; above all it must have high stability and be spin-proof. Its solidity should be in keeping with the stress imposed on the glider by gusts. It should be equipped with diving brakes to keep down the speed in vertical dives to a reasonable limit. These diving brakes have the additional advantage of stabilizing the glider about its longitudinal axis.

3.) The glider should be equipped with adequate blind-flying instruments. Besides the customary instruments, it should carry a turn-and-bank indicator and an artificial horizon.

4.) Cloud flying should be limited to occasions where the top of the cumulus is no higher than 3000 to 4000 m. During the summer, the dangers of icing and hail are thus avoided.

5.) In the cloud updraft, the flow is remarkably calm. On the other hand, violent turbulence is encountered at the transition from updraft to downdraft, especially at the rim of the cloud.

6. Inversions.

Inversions in the temperature zones.

Inversions are frequent during the summer, when the sky is filled with flattened isolated cumuli. Dry-unstable equilibrium prevails in the zone near the ground because of strong insolation, while the atmosphere is humid-unstable at the altitude of flattened cumuli. The slight vertical dimension of cumuli shows that there is a strong temperature inversion at higher altitudes that counteracts cloud updrafts. Figure 48 shows the corresponding temperature diagram: dry-unstable equilibrium ($\Delta T/\Delta H(\times 100) = 1.5°$) followed by indifferent equilibrium ($\Delta T/\Delta H(\times 100) = 1°$) up to 1900 m. At 1900 m the level of condensation and the base of the cumuli is reached, and equilibrium is thus humid-unstable ($\Delta T/\Delta H(\times 100) = 0.45°$). This state extends only throughout a thin layer. At 2200 m the atmosphere is completely stabilized by inversion of 2.3° in 350 m. In this example, the vertical speed can be computed by means of the differences of the dry and humid adiabatic states and the curve describing the state of the air. In the lower unstable layer, vertical speed increases from 0.1 m/sec at ground level to 1.4 m/sec at 1500 m altitude. Between 1500 and 1900 m (indifferent equilibrium) it remains constant at 1.4 m/sec. In the humid-unstable layer, between 1900 and 2200 m, the speed increases rapidly from 1.4 to 4.7 m/sec. Higher up, the air mass exceeds its equilibrium position and sinks from 2600 back to 2200 m at -2.6 m/sec.

This is a characteristic picture of summer high-pressure conditions in the temperate zone. Although diurnal temperatures are quite high, the possibility of gliding is limited to altitudes between 2500 and 3000 m. However, between ground and inversion layer, the upwind conditions for gliding are excellent.

Inversions in tropical and subtropical regions.

In tropical and subtropical zones, the situation described above persists throughout whole seasons. Over oceans and trade wind areas located between the 10th and 35th parallels and over the corresponding land masses, the atmosphere is characterized by its stationary character, distinguished by permanent or at least seasonal inversions at altitudes between 2000 and 3000 m.

Despite the considerable heating of the lower layers, the rising air masses are blocked at medium altitudes by this inversion, so that great clouds cannot form. This explains the constant aridity of the deserts which encircle the globe between the 20th and 30th parallels and the existence of dry seasons in tropical and subtropical zones. It is incorrect to assume that the tropics and subtropics are exceptionally well suited for gliding merely because of high temperatures prevailing in these regions.

It is rather the degree of stability of the atmosphere which must be considered. While the lower layers of these zones are very unstable and very favorable for long-distance glider flights during the dry season, one cannot hope to attain altitudes above 3000 m. However, during the rainy season the instability extends to high altitudes and thermal updrafts may go as high as the limit of the troposphere.

Gliding expedition to Libya.

The expedition to the Libyan desert in the spring of 1939 fully confirmed the above statements on the limitation of the altitudes attainable due to the effect of the permanent inversion.

On April 2, a measuring flight was carried out at Homs. It showed the inversion to be at an altitude of 1500 m. The four gliders which took off that day all reached ceiling altitudes between 1300 and 1800 m (Figure 39). Three measuring flights at Garian on April 12 showed that the inversion layer was between 2200 and 2600 m. The gliders which took off that day all reached the same ceiling altitude. The map of long-distance flights (Figure 49) from Homs on the coast or from Garian in the interior shows that remarkably great distances could be covered in these flights, despite the comparatively low flight altitudes and the routes necessitated by desert conditions.

7. Measurement of the field of vertical flow

The ability of gliders to move about in the atmosphere permits us to determine the field of vertical movements in a zone which extends up to cumulus clouds, thus giving us an excellent over-all view of the actual structure of thermal movements. These measurements feasible only in gliders are striking proof of the many possibilities in the utilization of gliders for aerological research.

It will be even more interesting to determine the same field inside clouds, not only for gliding, but also for all fields of aeronautics.

For these measurements a quadruple optical recorder was used, which records dynamic pressure, air pressure and vertical acceleration. Figure 50 shows a section of one of these recordings. The altitude curve was derived from the air pressure curve; and the curve of true air speed of the "Kranich" glider, which was used in the tests, was taken from the curve of dynamic pressure. Since turbulence causes slight fluctuations in dynamic pressure, mean values are used. The speed and duration of flight furnished the relative distance covered. The sinking speed of the glider was obtained from its "speed polar" (Figure 51). This with the vertical speed of the glider shown on the recording gave the vertical component of the speed of the atmosphere. The acceleration meter consisted of a inertia block in double-T shape connected to two pens. It had a measuring range of 1 to 4 g's. Figure 52 shows a sketch of a measuring flight.

The following figures give the results of measurement:

a) Figure 53 shows a simple regular vertical movement. On the day of the recording, an east wind prevailed of a velocity of 2 m/sec on the ground and 5 m/sec at higher altitude. Temperature gradient was 0.9°C; that is, nearly indifferent equilibrium. Vertical movements were blocked at 2200 m by an inversion. The base of the cumulus was between 2000 and 2200 m. The updraft field extended from 900 m to 2100 m; no measurements were performed below an altitude of 900 m. The thermal "chimney", to use a particularly appropriate expression, became wider, growing from 1 km diameter at an altitude of 900 m to 2 km at 2000 m. The inner core of the upwind field, with a vertical velocity of 2 to 3 m/sec, was narrow up to approximately 1300 m (300 m diameter) and then widened to a diameter of 1500 m.

The ascending velocities were remarkably regular throughout the field, ranging from 0.5 to 1 m/sec at the outer edge to 2 to 2.5 m/sec in the inner core. This is due to the indifferent equilibrium of these air layers, which does not impart any acceleration to the rising air masses. Since only small distances were covered in the downwind field, its exact extent could not be measured; but in all cases the downward velocities were less than the upward velocities. However, in two isolated cores downdrafts of a velocity of 2 to 2.5 m/sec were found. Above 1600 m, at the limit of the measured region, new upwind fields were reached. This cell-like arrangement of vertical movements should correspond to a regular exchange pattern of the air in a vertical sense, a hypothesis which is used, among other things, to explain cloud lanes. Likewise, the vertical distribution of temperatures, which represents a discontinuous surface caused by an inversion, also fits these indications.

b) Figure 54 represents results of measurements of quite another situation: The wind on the ground was a W to NW wind of a velocity between 2 and 4 m/sec. The temperature gradient was 0.87°; that is, humid-unstable. Thus, heavy billowing cumulus clouds developed. The cumulus underneath which measurements were performed had a diameter of 6 km at the base and the updraft field was 4.5 km; nevertheless, it narrowed down toward lower altitudes and was interrupted at 1400 m. That meant that the updraft field, as well as the cumulus cloud, had already passed the stage of their maximum development and were disintegrating. Despite this fact, the updrafts in the core were still quite strong in the center and within a 700 m diameter; the air had an upward velocity of 4 to 6 m/sec. This updraft field had several special features:

1) The horizontal gradients of upward velocities were greater on the windward side of the cumulus (East) than on the leeward side. At the windward side the upward velocities increased from 0 to 4 m/sec over an average distance of 700 m; and at 1900 m this gradient reached 7 m/sec over a 500 m distance. On the other hand, on the leeward side the vertical velocity decreased over a distance of 4 km from +4 m/sec to 0. From this can be concluded that pilots should expect to find the strongest updrafts underneath the windward half of clouds. At the transition from windward to leeward side, violent gusts are found, which are due to the higher horizontal gradient of the vertical velocity.

2) Measurements showed that within a cloud and even below it downdrafts were found on the windward side. As previously stated, these downdrafts were much more feeble than updrafts.

Although the field just described was already cut off at the bottom and disintegrating, there was a large updraft zone on the windward side. During the measuring flight, it was found to reach an altitude of 1500 to 1600 m. It formed a regular chimney extending for 1 to 1.5 km, with constant upwinds of 2 to 3 m/sec and a core with +5 m/sec at 1400 m altitude. Once the updraft reached the condensation level, it gave rise to the formation of a large cumulus right next to the one which was disintegrating.

c) Figure 55 is the third example of vertical movement of thermal fields of 23 July 1937 and represents the results of measurements of much more irregular conditions than the two previous figures. It shows updrafts and downdrafts of very irregular distribution and choppy character. The downdraft zone contained many cores in which the downward velocity was as high as the maximum upward velocity. At an altitude of 1750 m, there was a small core with a velocity of -3 m/sec, right next to 1 m/sec updrafts. At an altitude of 950 m, downdrafts of 2 to 3 m/sec were right next to updrafts updrafts of 3 to 6 m/sec. The same abrupt transitions were found between 500 and 900 m to the right of the measured zone.

These measurements are an example of thermal emissions pulsating in rapid succession above a fixed releasing point (forest edge). The emissions occurred every 12 to 15 minutes. The updraft core of 2.5 m/sec at 1800m, which was in the process of disintegration, was followed by a new emission between 500 and 1300 m which climbed rapidly at a speed of 3 m/sec and reached speeds of 4 to 6 m/sec in its core between 700 and 1000 m. A third emission was rising on the left, at 2 to 3 m/sec, which had reached an altitude of 900 m at the instant of measurement. These moving air masses were replaced by downdrafts, which returned with more or less the same speed to the layers close to the ground. This was a strong thermal mixture with great activity of the thermal boundary layer on the ground. While the general temperature gradient of the day was not very great and the atmosphere was in an indifferent state of equilibrium (0.98°C per 100 m), the strong insolation created intense overheating of the thermal boundary layer, resulting in the periodic emission of "air bubbles".

The interplay of updrafts and downdrafts, shown in Figure 23 for this boundary layer, was repeated here in the same manner, but on a much larger scale.

8. Cloud lanes:

Quite often one can observe rows of cumuli in the sky, lined up in the direction of the wind and sometimes forming along regular ribbons. This cloud arrangement has been given the appropriatedterm "cloud lanes" by glider pilots.

These lanes are particularly well formed over flatlands or sea and suggest the hypothesis of some well-ordered convection process resembling certain known flow phenomena of unstable liquids ~~known~~. The convection of flow of a thin unstable liquid layer has often been studied experimentally in physics.

Analogy from Physics.

As early as 1901 Henri Benard exposed the analogy between the cellular structure of unstable liquids and the organization of rows of clouds. Idrac showed that air flow between two plates, with the lower heated, takes on the aspect of vortices whose axes are parallel to the flow.

Figure 56 shows the structure of an immobile and unstable liquid. The cellular polygonic honeycomb-like arrangement can be distinguished. The liquid rises in the center of the cells and descends along the edges. Under the influence of a horizontal flow added to it, the cellular vortices are transformed into rows lined up parallel to the direction of flow. Figure 57 shows the regularity of this arrangement. Figures 58a and 58b taken from a treatise by Dr. Avsec also show the ribbon-like arrangement of the vortices.

The pattern of cloud lanes.

If we want to transfer this phenomenon to the atmosphere, we must assume the existence of an unstable layer of air between two stable layers. In calm air, cellular clouds, isolated clouds in a row, regularly spaced cloud fields, altocumuli or stratocumuli may form in this layer. Under the influence of wind, the same atmospheric conditions will give rise to the formation of cloud lanes; a discontinuity exists in the wind velocity between the unstable and the upper stable layer whereby friction between the two layers is increased, but there is little variation in wind direction.

These statements have been confirmed by aerological measurements. Dr. Sobhagmal in "Beiträge", Vol. 17, 1931, has given the temperature gradients and wind discontinuties:

Below cellular clouds	0.54°C per 100 m
Interior	1.40°C per 100 m
Above	0.59°C per 100 m

He noted wind variations from 7 to 12 m/sec at the transition between unstable layer and stable layer.

Figure 58 shows the horizontal flow lines. Along the line of convergence is found an updraft, while a downdraft lies along the line of divergence. Figure 58b shows a vertical section of the field, which represents counter-rotating vortices. Figures 58c shows the helical air flow and the formation of cumuli.

Different types of cloud lanes.

a) In the tropical seas and trade wind zone, cloud lanes are particularly striking, since optimum conditions exist there: the regularity of the trade wind, the stationary temperature conditions of the tropical latitudes with an inversion at a proper altitude, and the homogenous surface of the ocean. Figure 59 shows one of those cloud lanes which the author found at 5° S latitude in the Atlantic. It extended clear across the sky and completely cut it in two.

b) Frequently, isolated mountain massifs give rise to the formation of a cloud lane when they penetrate a discontinuity surface forming the upper limit of an unstable layer. Kampe de Feriet studied one of these marginal vortices on Mt. Cervin and drew from it Figures 60a and 60b. In this humid-unstable zone, the row of vortices is made visible by a long lane of clouds originating on a rocky pinnacle. Figure 61 shows a clearly distinguishable cloud lane with the origin of the cloud formations visible in the picture. This photograph was taken during an East wind, which is frequent in spring and fall. The atmospheric conditions here are exactly those of the thermal structure analyzed above.

In our latitudes, too, cloud lanes can be observed in hot humid subtropical air, whose altitude is limited by air masses in the process of disintegration.

We shall return to the subject of cloud lanes in connection with the discussion of atmospheric waves.

The use of cloud lanes for gliding:

Again it was gliding which attracted attention to this formation of longitudinal vortices. These cloud lanes are of great interest to the glider pilot, since they provide a continuous updraft field in the direction of the wind and thus afford ideal conditions for long-distance flights.

Ordinarily, the distribution of thermo-convective updraft zones is arbitrary. In long-distance flights, the pilot must look everywhere for clouds which will permit him to gain sufficient altitude by spiraling to continue his flight, and these spirals will reduce the relative speed of his flight. The barogram of such a flight consists of an irregular series of ascents and descents, while the barogram of a flight underneath a cloud lane looks altogether different. Underneath cloud lanes, there is no need for flying spirals, for flight proceeds in a continuous updraft zone at nearly constant altitude. Therefore the effective speed of such a flight, performed without spiraling and with tailwind, may be as high as 70 to 80 km/hr (Figure 62).

In order to break present distance records, high average speed is required. In mid-summer it is possible to perform thermal flights of 8 hours duration. In normal weather and at an average speed of 30 to 40 km/hr, 250 to 300 km can thus be covered, while an average speed of 60 to 70 km/hr is required for flying a distance of 500 km or more. Such speeds are easily attainable in flight under cloud lanes.

9. Flight in weather fronts:

The first flights covering more than 200 km distance were flights in weather fronts. It is an obvious thought to use the updraft formed by rising warm air ahead of a cold front and thus to "ride along" with the front. Since cold fronts travel at speeds of 40 to 60 km/hr, considerably long flights can be carried out by this method.

Description of a front:

Figure 63 gives a schematic representation of a front. The passage of a cold front is accompanied by a sudden wind shift. A characteristic cumulus front, which frequently starts with a line squall and ends with a heavy shower, causes the warm air to rise quickly. The wind suddenly becomes stronger and shifts its direction abruptly, at the same time the temperature drops considerably and the air pressure rises gradually. Ahead of the front, the vortex motion of the air is characterized by a roll of clouds.

Utilization of a front for gliding:

Flying in the zone just discussed would be risky. On the other hand, ahead of the front and outside of this cloud zone is a completely calm upwind field. The first flight of this kind took place in July 1929 (Figure 64). It was most informative. The airplane took off two minutes before the arrival of the front. The airplane "hitched itself" to the upwind zone of the front after 26 minutes of flight and then climbed steadily from 1300 to 2900 m. The greatest climbing speed, 4 m/sec, was achieved between 1400 and 1800 m. The barogram shows that once the ceiling was reached the flight continued at nearly constant altitude, between 2100 and 2400 m, for nearly two hours. This leads to the assumption that an extremely large and regular upwind field must exist. According to the pilot's statements, he flew 5 to 10 km ahead of the front and underneath the alto-cumuli separated from the front by a strip of blue sky. Figure 65 shows the barograms of two front flights carried out later (July 1931).

These examples of front flights indicate that the updraft fields ahead of cold fronts all have the same structure and can be considered as quasi-stationary. Below we will discuss further the conclusions which can be drawn from this.

At this point let us simply state that front flights are not dangerous as long as they are carried out in a cloud-free zone ahead of the front and as long as the pilot does not let the airplane be carried away from the front and, above all, does n't let it be drawn into the cloud zone.

In carrying out a front flight, the glider should preferably be towed by a powered airplane to a point ahead of the front rather than start from the ground, since the instant of the wind shift, when the upward gusts set in, is always difficult to catch.

10. Thunderstorm flights.

In the examples discussed up to now, ascending masses air were held back by elevated stable layers, so that updrafts could hardly go above 4000 m. These updrafts can be used without danger by a glider equipped with the usual blind-flying instruments.

Now, however, we shall take up the case where the unstable equilibrium of the atmosphere allows the air masses to rise through the entire troposphere. This occurs in the formation of large cumulo-nimbus clouds, which bring about violent local heat thunderstorms.

For our findings on updraft conditions, and on the dangers of turbulences, icing, hail, and cold occurring in thunderstorm clouds, we are indebted to glider pilots who have risked flying through these clouds to reach the limit of the troposphere.

A pilot's report on a thunderstorm flight:

"The cloud reached an altitude of 5500 m. We intended to fly through the border zone of the cloud and entered it at a speed of about 130 km/h. The airplane was drawn into the cloud's interior by the turbulence. This turbulence soon stopped; while it did not actually get dark, the cloud thickened to such a degree that it was hardly possible to distinguish the wing tips. Besides, fine ice crystals caused sharp pains in the face, making it difficult to look out of the cockpit. At 3100 m the goggles and instruments and finally the entire airplane became covered with ice; an opaque 6-mm piece of ice was removed from the control stick. Soon after, around 3000 m, hail began to fall, with the largest hailstones reaching the size of cherries. At 2400 m the hail stopped and turned to rain. We emerged from the cloud at 1300 m."

All through the descent, the engine was throttled; nevertheless the airplane maintained constant altitude for 20 sec at 3000 m, and at 2400 m it climbed more than 100 m within 40 sec. The sinking speed of the airplane was around 9 m/sec, so that the updrafts reached a velocity of 8 m/sec at 3000 m and 12 m/sec at 2400 m.

Investigation of the special features of thunderstorm flights.

The observations made in gliders flying through thunderstorms are even more striking. Figure 66 represents the temperature-altitude diagram of an actual flight. The humid-unstable difference, i.e. the difference in temperature between the humid adiabatic and the vertically still air, is very great. From 2200 m to the tip of the cloud, the temperature difference increases just as the upward velocity in the cloud continues to increase.

The upward velocities have approximately the following distribution: Underneath the base of the cloud it is 1 to 3 m/sec. Starting at the level of condensation the velocity increases rapidly, rising to 5 to 6 m/sec within about 100 m. This velocity continues to increase with increasing altitude. Barograms indicate average velocities which may exceed 20 m/sec. An analysis of a set of barograms yielded average speeds of 10.13, 17, 20, and 27 m/sec in various layers at altitudes of several 1000 m. Velocities of 20 m/sec and above are generally restricted to altitudes extending from 3000 to 5000 m.

Barograms like the one in Figure 67 indicate a continuous climb throughout several 1000 m and reveal the existence of a completely non-turbulent and regular ascending flow. Most flight reports speak of a "regular climb in an updraft of astonishing calmness"; or they state: "No turbulence during climbing". In the lower portion of the cloud the updraft field is wide, but narrows at altitudes of great vertical velocities, so that it is sometimes necessary to fly in very tight spirals.

In some barograms it can be seen that the climb has been interrupted very suddenly by a downdraft of 4 to 6 m/sec velocity. At the transition from updraft zone to downdraft zone is a narrow turbulent area of unimaginable violence, so that pilots have great difficulties and several gliders have crashed because of it.

However, the main dangers are hail, cold, and icing. The icing is due mostly to the cooling below 0°C of water droplets carried up to high altitudes. Its most dangerous effect is the blocking of controls and diving brakes and the icing up of the Pitot tube.

Hail is encountered between 2600 and 7000 m. In several cases, it has had a disastrous effect. The wing fabric, fuselage skin, and cowling have been completely riddled.

Obviously, proper equipment is required for the cold prevailing at high altitudes. However, the pilot usually does not foresee the altitude which he will reach and therefore fails to equip himself properly. It is most unpleasant, in an airplane not hermetically sealed, to find oneself in a thunderhead at 7000 m altitude and a temperature of -25°C while it is + 30°C on the ground - and this is quite likely to happen. Furthermore, the pilot might be forced to bail out at this altitude, becomes exposed to the open air and risks freezing to death. In one case, a pilot who had parachuted from his airplane was exposed to these low temperatures for a long time, because the updraft was so powerful that his parachute carried him upward.

Glider flight in a thunderstorm:

On 12 July 1947, the Swedish pilot Axel Persson succeeded in achieving a climb of 8200 m in a thunderhead (Figure 67). The instability was very great, and the cumulus extended through the entire troposphere. The theoretically determined upward velocity exceeded 20 m/sec, and the glider although loaded down with ice climbed at a rate of 15 m/sec. The icing was not very heavy, since the danger zone, between 0° and -10°C and from 2500 to 3900 m, was traversed within two or three minutes. The pilot also stated that the intensity of the icing of his airplane was reduced by the application of glycol to its skin. His report does not mention any other difficulties such as gusts or hail, so that one is led to believe that the pilot remained in the zone ahead of the front, which is much less dangerous than the center or the rear portions.

Nevertheless, there is no point in coming out in favor of such thunderstorm flights, because they are fraught with grave dangers despite this reassuring report. When imagining the situation of a glider, with its surface iced up, the blind-flying instruments frozen, essential components seriously damaged by hail, and, while in this state, also subjected to the most violent gusts, one feels constrained to give serious warning to all glider pilots who do not have special airplanes and equipment for such flights.

Structure of a thunderstorm cumulus:

Now that we know some of the features of flight through thunderstorms, we can examine the structure and the development of thunderstorm cumuli.

a) The life of the cloud:

During the first stage of the development, updrafts predominate in the interior of the cloud. From the point of view of vertical flow, temperature distribution, and the phases of the cloud, there is symmetry. At that moment there is no rain as yet, since the updraft carries the condensed water droplets along with it.

When the isotherm of 0° is passed, the water drops do not change their state, but continue to exist in the supercooled state. Icing occurs in this zone. Then the fully developed cumulus passes the -12° isotherme, and at this altitude its structure changes and the development characteristic of thunderstorms sets in. The liquid droplets freeze at -12° and form crystals in the upper portion of the cumulus. The exact temperature of freezing is not known, but the appearance of ice in clouds below a temperature of -12° has been established. The vapor is highly supersaturated with respect to the solid state and the ice crystals quickly grow in size, while the updraft becomes weaker. Thus the tendency toward precipitation increases, the closer the altitude of thermal equilibrium is approached. Finally, the velocity of fall of the crystals becomes greater than the upward component of the wind and the crystals therefore begin to fall. They liquefy in the lower layers and reach the ground in the form of large raindrops.

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When rain starts, the discharge proper of the thunderstorm has begun. The initial symmetry disappears and a front and train are formed. The crystals and water drops falling from higher altitudes are colder than the surrounding air and cool it. The temperature distribution becomes asymmetric, with a warm front and a cold rear portion, as is shown by the 0 and -12° isotherms in Figure 68. The air which is cooled in the zone of precipitation will thus also descend, on the one hand because its density has increased and on the other hand because it is carried along by the precipitation. A strong downdraft zone is formed next to the updraft zone, and the air flow of the cumulus is transformed into a great whirl which is maintained by the difference in temperature of points on the same level; namely points ahead of the front in the warm rising air and the other points behind the front in the cool descending air. These two air masses of different temperature form a cold pseudo-front on the ground. Its limit is the curtain of rain marked in the air by the squall cloud and the roll of clouds underneath the base of the cumulus, produced by the lowering of the level of condensation in the cold air.

b) Allied phenomena: We can now also explain other physical phenomena which take place during the principal phase of a thunderstorm.

1) The turbulent zone: The strongest gusts will be found at the border between updraft and downdraft, approximately in the center of the cloud (Figure 68). As the life of the cloud progresses, this region of gusts moves forward and makes the updraft zone narrower.

2) Hail: Hail is another dangerous phenomenon of the thunderstorm. The crystals in contact with supercooled water are covered by a layer of ice. After multiple collisions these crystals turn to hailstones. Hail apparently forms when a crystalline nucleus traverses the supercooled zone of the cumulus several times, i.e. it must travel several times through the zone of updrafts and downdrafts. This is quite possible if the hail zone is located in the center of the cloud, astride the boundary layer discussed above, and above the cold pseudo-front.

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3) Electrical charges. The electrical charge is a unique characteristic of thunderstorms. The problem of electricity in thunderstorms has not yet been completely solved. There are two principal theories: Wilson's theory, which is based on the capture of ions by the elements of the cloud; and Simpson's theory according to which the electrical charge of the thunderstorm is due to the bursting of the water drops, which allows the charges to separate according to the Lenard effect.

According to the first theory, positive ions are captured by the rising drops which are thus charged positively; and falling drops capture negative ions, thus imparting a negative charge to the lower portion of the cloud.

However, according to Simpson it is the Lenard effect which predominates. The water drops become positively charged after displacement, while the surrounding air is negatively charged and the initial drops are neutral. The drops are stable as long as their radius does not exceed 0.225 mm. Above that dimension they deform and break up. Large drops of rain are found in the downdraft zone below the 0° isotherm. They split and transmit their positive charges to the ground while the negative charges accumulate in the cloud below the 0° isotherm. There are also large drops in the updraft zone; they also disintegrate, the positive droplets remaining in the updraft, with the negative charges accumulating in the upper portions of the cloud (Figure 68).

Concluding remarks on thunderstorm flights:

The above discussion may furnish pilots with some instructions, but it should not under any circumstances be considered an invitation to thunderstorm flying. These instructions are meant only as safety rules in case of an unexpected encounter with a thunderhead. Furthermore, the theories which have just been discussed are not yet certain by any means, and the deductions made should be accepted only cautiously.

On the outside of the cloud, the rain front is an indicator of the non-symmetry of the cumulus. The front is the least dangerous zone of the thunderstorm; it contains a regular updraft and there is little chance of running into violent gusts or hail. It is probable that Persson achieved his altitude record in this zone. The friction layer between the updraft and the downdraft zones, located approximately at the center of the cloud during the principal phase of the thunderstorm, is the most dangerous zone, because of very violent gusts and hail there.

- END OF CHAPTER 3 -

CHAPTER 4

WAVE FLIGHT

A. Historical.

The first wave flight was carried out in the Giant Mountains of Silesia in 1933. What was startling then was the fact that the flight took place on the lee side of the mountain during a foehn wind; but according to the then-known dynamics of foehn winds, this warm dry wind could produce only downdrafts on the leeward slope, so that it was difficult to explain the existence of updrafts on that side that allowed gliders to reach high altitudes.

Since 1933 a great number of flights in foehn wind have been carried out and over all kinds of mountains, som flights reaching altitudes of 8000 m, and even 11400 m over the Alps. These results allow us to form a judgment of this new possibility of flight, and they indicate such a remarkable widening of the field of glider flight that we can speak of an entirely new development.

The first phase of glider flight began with the solution of the problem of soaring over slopes in 1922, when a distance of 100 km and an altitude of 375 m above the starting point was reached.

The second phase began in 1928 with thermal flight. The best performances registered by this method of flight are a distance of 750 km and a gain of altitde of 8200 m. Glider flight proved itself to be no longer limited to mountains, but ventured out into the plains and took its permanent place in the field of aeronautics.

At the present time we cannot yet predict the possibilities of the third method, namely wave flight, but the research carried out and the results obtained prove that wave flight constitutes an important turning point and that the stratosphere has most likely been opened to gliders.

B. Free atmospheric waves - Gravitation waves.

In order to understand the principle of wave flight, we must first discuss free waves.

1. Analogy from Hydraulics

Since liquid waves are well known, we shall use the analogy from hydraulics. These waves originate at the surface of the water (namely, at the surface of separation of two different media, water and air). The molecules of the two media are driven from their position of equilibrium by an external influence (in our case by the wind) and start oscillations around their position of equilibrium under the influence of gravity. These oscillations are transmitted to neighboring particles and cause a wave motion which propagates on the free surface. The speed of propagation of these gravitational waves depends on wavelength, density of the two media, and depth.

As mentioned previously, there exist in the atmosphere inversion layers which violate the normal principle of decreasing temperature with increasing altitude, for in that region air temperatures increase instead. These inversions form limiting surfaces within the unlimited atmosphere; thus there is a density discontinuity across the section, so that, just as on the surface of the water, waves can form due to the gravitational field under the influence of wind. These waves are mobile and are perpendicular to the direction of the relative wind of the two flows. The wavelength depends on the discontinuity of both density and temperature and on the velocity within the inversion. Long waves are created by a great discontinuity of velocity, and short waves by a great discontinuity of density or temperature. Long waves propagate faster than short waves.

FigurFigure 69 shows a representation of the flow lines of a wave. On the front side of the wave in the direction of its propagation, the motion is an ascending one.

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2. Cloud formations.

These waves can be made "visable" through the formation of so-called "wave clouds". If the atmosphere below the inversion is almost saturated with water vapor, then the rising part of the wave will lift sufficiently to initiate condensation, while the wave will disintegrate in the descending part of the wave. Thus regularly spaced cloud banks are formed which make these atmospheric waves visible.

3. Experimental Study of Waves

a) Measurements by airplane

Let us assume a powered airplane flying directly in a boundary layer. In order to maintain the airplane at the same altitude, the climb of the airplane in the wave is compensated for by throttling of the engine, while the dynamic pressure remains the same and the intake is even increased. If the movement of the throttle handle is recorded, a qualitative measure of the movements of the air is obtained. Figure 70 is the diagram of a flight carried out on 17 July 1936 in an inversion layer, which shows: a) humidity, b) temperature, c) speed (dynamic pressure), d) outside pressure, e) vertical velocities (indirectly). The temperature variations in this recording give a particularly good picture of the wave motion.

b) Use of balanced balloons

Likewise, the use of statically balanced balloons allows one to determine the paths of waves at the inversion level.. Two examples are shown in Figures 71 and 72. The balloons were released in the inversion layer from an airplane. The paths show waves 2200 m long on 2 April 1930, and 1000 m long on 3 April 1930.

The mean vertical velocities are low, ranging from 0.6 to 0.8 m/sec.

c) Glider flights

Figure 3 shows the barogram of a long-distance glider flight of 165 km performed in 1931. The flight was carried out on a cloudless day, thus under critical thermal conditions. The air was dry-unstable up to the level of a slight inversion at an altitude of 1800 m. Above this inversion, the air was humid-unstable.

The barogram shows the course of the flight starting at its 68th minute. The curve H(t) shows a very regular succession of rising and falling air movements. These oscillations, from 800 to 1200 m, show a wave motion with regularly alternating rising and falling air, allowing the pilot to fly in a continuous straight line without having to either look for other updraft fields or fly in spirals. This characteristic indisputably proves that this flight was carried out chiefly under the influence of a wave motion located on a surface of discontinuity at an altitude of 1800 m.

4. The possible applications of free waves.

This example shows that these waves do have a certain importance in glider flight, but that one should not pin very high hopes on the applications of these waves. As a matter of fact, while there are no relative measurements of the damping of these waves with vertical distance, the vertical motion of these waves is probably not very strong.

At any rate, the above examples do show that the free waves propagating from the inversion layer are worthy of some consideration, although they can be utilized only occasionally in the course of a flight when the formation of wave clouds indicates these waves and fixes their position.

It should also be mentioned that cloud lanes can sometimes occur together with free atmospheric waves.

If wind directions above and below are fairly similar, (Figure 74), the crests of the waves will be more or less parallel to the lower flow and condensation at these crests will produce cloud lanes.

C. Stationary waves.

1. Differences between free and stationary waves.

The oscillatory motions which have led to the great accomplishements of wave flight are altogerher different from the gravitational waves discussed in Section IV, 1.

The upwind of waves known in gliding, which extends to a height of several km and is accompanied by the oscillation of large atmospheric layers, is such that the vertical motion is not damped as in the case of gravitational waves.

Furthermore, in contrast to free waves which propagate at the surface of discontinuity, stationary waves can attain considerable amplitudes. Such oscillations are initiated by the "basic waves" started by disturbances in flow particularly by those of orographic nature. If in this case flow velocity equals the velocity of wave propagation, then the wave will be stationary with respect to the terrain obstacle which caused it to develop. The amplitudes of these stationary waves may continue to increase as long as the oscillation generated by the obstacle continues to receive new impulses. The velocity of propagation of the wave itself depends on wavelength and on the depth and density of the oscillating layer.

2. Analogy from hydraulics

The phenomenon and theory of these generated stationary waves were first studied in the case of water flow. If water flows over an uneveness in the bed, the surface configuration of the liquid will differ according to whether flow velocity is greater or less than the propagation velocity of the basic wave created by the obstacle. The forms of water surfaces of separation are shown in Figure 75.

Figure 75a shows a "stream flow": The flow velocity is less than the velocity of the basic wave. In this case the surface of separation is lowered over the top of the obstacle.

Figure 75b on the other hand shows a great rise in water level, the height of the rise being greater than that of the obstacle itself, with the flow faster than the velocity of the basic wave. This is a "torrential flow".

Figure 75c shows a "stream flow" upstream from the obstacle and a "torrential flow" after the drop in the water level.

Figure 76 again demonstrates that the vertical accelerations connected with the variations in ground level cause stationary waves downstream from the obstacle.

The above statements apply to the free upper surface of a fluid. Taking up now the case of the atmosphere cut by surfaces of internal discontinuity (inversion layers) we can study the waves produced between two fluid media having the same velocity of flow but different densities. A. Defant has carried out experimental studies on this subject (cf. Defant: "Réflexions théoretiques et recherches experimentales sur la formation des cyclones et anticyclones en altitude") and determined the following results: If an object is placed in a canal filled with two fluids of different densities and flowing with a certain velocity, the surface of separation between the two fluids will react to each movement of the obstacle with the formation of wave disturbances. The wave disturbance reaches its greatest amplitude at a certain speed (in one particular experiment this speed was ~~experiment~~ 8 cm/sec, when the difference in densities was 0.015 at 15°C, depth of the fluids 20 cm and 5 cm, and the model 20 cm long and 5 cm high). The form of the flow in the canal corresponds exactly to that of a free surface of a fluid which passes from stream flow to torrential flow above an obstacle. (Figure 75c).

However, while in a homogenous fluid with a free surface the disturbance effect requires high speeds, in the above experiment small flow velocities sufficed to create noticeable deformations in the surface of separation. This holds true when the densities of the two superposed fluids are very similar and the fluids respond to weak impluses created by an obstacle.

These fixed wave disturbances of incompressible fluid media will also occur in a compressible medium such as the atmosphere. It is possible to go from this limited case of fluid flow to the case of the expansible atmosphere, by considering the differences in densities and velocities according to the principle of similarity as studied in dimensional analysis. Otherwise, nothing in the principle of wave motion is changed by the compressibility of the atmosphere.

The above can be summarized as follows:

1. An air mass of sufficient dimensions flowing over an obstacle takes on a wave motion above the leeward slope of the obstacle without having been supplied any external energy.

2. The velocity of propagation of these waves depends on wavelength, temperature distribution with altitude, and dimensions of the moving air layers.

3. When the wind velocity approaches the velocity of propagation of the waves, the amplitude becomes great; in this case, the waves become stationary with respect to the obstacle and the wave motion caused by the obstacle is maintained constantly.

4. Depending on whether the wind velocity is greater or smaller than the velocity of propagation of the waves, the flow above the top of the obstacle will by upward or downward.

3. Mathematical solution of the problem of stationary waves.

The problem of wave disturbances above terrestrial obstacles has recently been treated by Lyra on an entirely theoretical basis (Cf. Zeitschrift fur angewandte Mathematik und Mechanik, March 1943). Starting with fundamental thermo- and hydrodynamic equations, Lyra calculated the magnitude of the disturbances of an initially homogenous flow. Assuming a stable atmosphere, he established an equation for waves which allows one to represent the upwind field to great altitudes. The upward velocities of such a wave disturbance are shown in Figure 77a.

Directly above the obstacle, we find a slope upwind, which at higher altitudes changes into a downwind in the zone where the flow lines disintegrate. The first of the leeward waves develops strongly, starting at an altitude of 2 km, and its amplitude increases with altitude. The second wave is already strongly damped and is perceptible only at high altitude. The third wave, however, is hardly noticeable at all. As Figure 77b shows, a terrain obstacle which ends in an elevated plateau suffices for the generation of waves. In this case the updrafts of the first wave are particularly strong.

4. Experimental study of stationary waves.

The study of air flow in mountains began with the beginning of glider flying during the period of slope soaring; after this period attention turned to the wave motion of flow near obstacles.

The first measurement by means of statically balanced balloons was performed in 1924. It showed the existence of waves behind mountain ridges (Figure 78). Actually, this measurement considered all by itself did not allow any definite statement as to whether the flow was that of a wave or turbulent field, but according to the present state of our knowledge it is a leeward stationary wave.

a) The simple regular form of coastal dunes over which ~~even~~ steady sea winds blow are particularly suitable for the carrying out of such measurements. A series of measurements was carried out in 1928. They clearly showed the existence of stationary waves to the leeward, as pictured in Figure 79a, b. The ballons were released on June 6 1928, between 1615 and 1707 hrs. The interval of time between the various measurements was long enough so that the flight paths could be considered as quasi-stationary. The flight paths of the balloons followed closely the waves. After their downward flight just behind the dune, they show a wave of fairly great length. According to the flight paths of balloons Nos. 42 and 43 (Figure 79a), the wavelength is approximately 3200 m for a wind of 10 to 11 m/sec.

Balloons Nos. 55 and 56 were observed for a wind of 10 m/sec. The wavelength varied between 3000 and 4000 m. Other measurements carried out four weeks later over the same terrain for a wind of 12 m/sec showed great resemblance to previous ones (wavelength 4500 m). The crest of the principal waves in all flight paths lies about 3000 m from the slope. These measures thus show quite clearly the existence of stationary waves.

b) The flight paths shown in Figures 80a, b were recorded in the mountains on the same day. The balloons were released from an airplane just below and above an inversion between 1400 and 1500 m altitude. There was a second inversion between 1800 and 2000 m altitude. The two flight paths (Figures 80a and 80b) are completely in agreement and show a lifting wave ahead and above the obstacle and a trough immediately behind it, and then a second crest further to the leeward with a wavelength between 3.5 and 4.5 km.

c) Figure 81 shows a particularly interesting picture, namely that of very clear waves of small length (300 to 400 m) for a wind of only 2 to 3 m/sec. Other measurements carried out over the same terrain two days earlier had shown analogous wave formations 360 to 1000 m long for a wind of 3 to 5 m/sec. The measurements of Figures 29 and 30 are also interesting. They show a characteristic wave form to the lee of the slope. The top of the second wave is in a small cumulus cloud. The wavelength is approximately 800 m for a wind velocity of 3 m/sec. The characteristics of these two measurements, made at an interval of 30 minutes, are completely analogous. The position of the upwind zones has not changed. A third measurement performed four hours later showed an identical picture.

Thus, stationary waves exist in a slight wind even for thermal instability conditions.

5. Foehn waves in the Giant Mountains.

The meteorologist J. Kuttner has studied in great detail the waves over Upper Silesia. His theoretical considerations have been completed and confirmed by a number of observations and flights by the author.

Figure 82 shows the scheme of Foehn waves on the leeward side of the Sudeten mountains. Furthermore, the observations made during flights have made it possible to reproduce a section of the flow, which shows clearly the structure and spatial position of the wave system (Figure 82). Above the lower turbulent layer, which is marked by fixed eddies, the regular wave flow extends up to about 6000 m and creates at high altitude the great Foehn cloud, the well-known "Moazagotl". Figre 83 gives the position of the upwind zones with respect to the obstacle. The first wave is located about one wavelength from the mountain, i.e. 7 to 8 km, so that the third wave is about 20 to 25 km distant from the mountain range.

The Silesian wave reaches its highest altitude at 8600 m.

As the studies of Kuttner have shown, the various cloud forms are quite typical. The high heavy Foehn cloud is so striking that it has its own name in the Silesian dialect, "Moazagotl". Figures 84 (16 Sept. 1937) and 85 show two typical examples of Foehn clouds with the convex lens form (lenticule) clearly discernible. This peculiar form is the result of wave flow, with one layer nearly saturated with vapor and lifted to the crest of the wave, where it is cooled to the condensation point. The upper limit of the cloud is determined by simultaneous inversion.

Figure 86 shows an excellent example of a roll of eddies, the lower portion of a wave flow, which formed during a northwest wind over the Swiss mountains between Neuchatel, Soleure and Olten. If the wave flow has come that close to the ground, it is certainly possible to fly in it after a start by winch.

Kuttner's findings can be summarized as follows:

a) When flying with tail wind and perpendicularly to the wave system, the glider passes gently and continuously from ascent to descent and vice versa, without the slightest turbulence or sudden transition.

b) When flying with headwind, the airplane whose air speed is nearly the same as that of the wind keeps climbing in the forward part of the wave without changing its position with respect to the ground. When flying parallel to the wave, the wave operates like a huge cloud lane with a constant updraft field.

c) In order to descend, the pilot has only to find the rear of a wave and is immediately carried to earth in a smooth flow.

Figure 87 shows the barogram of a wave flight over the Giant Mountains and shows very well this steady and regular climb in a constant upwind field.

General statements on waves of terrain rises

It is natural that until now most attention has been given to the Foehn waves in the mountains, because these waves permit the execution of spectacular flights, so that nobody has bothered much about the waves of terrain rises caused by medium-sized obstacles.

These waves correspond to the damming effect above an obstacle in a "stream flow" with a flow velocity greater than the velocity of propagation of the wave. They produce a great deviation in flow immediately above the obstacle (Figure 75b), of an altitude many times that of the obstacle itself.

Such waves have already been found over minor obstacles, rising 100 to 200 m with fairly steep slopes, in wind velocities of the order of 40 km/h, and have allowed flights up to 2000 m altitude. These waves are produced particularly during winter, when only slope soaring is possible. They have been utilized in a number of countries. In France, the regions of Corbas and Port-Saint-Vincent can be named.

6. Stationary waves over the Alps.

General statements

After the remarkable results obtained in the waves over Silesia, with elevations of 1200 and 1500 m of the terrain, it was to expected that wave flights over the Alps would particularly successful. With this intention, the German Glider Institute (Deutsche Forschungsanstalt für Segelflug) carried out a series of research flights over the Eastern Alps, between the Hohe Tauern (Grossglockner 3798 m) and the Upper Bavarian Highlands (500 m) in the region of the great Foehn cloud which is familiar to the inhabitants of that region.

In order to compare the flight results with the theoretical determinations of Lyra, Ruth Staffin systematically computed by that method the vertical velocities of the air ~~in the northsouth~~ North-South section of the Hohe Tauern range (Figure 89). Above the Northern portion of the Tauern range (from the Grossglockner to the Kitzsteinhorn) prevails a strong downwind between 3000 and 6000 m. Then, 20 km to the leeward of the main range, follows an updraft which increases with altitude, shifts toward the South and then decreases rapidly above 10 km. On the other hand, the wave on the windward side is weak, becomes stronger with the altitude, and shifts toward the south starting at the level of the stratosphere. Despite the inevitable simplification caused by a schematic calculation, the results agree well with these of flights.

Description of the flight of 11 October 1940..

The most convincing example of the great possibilities of wave flight over the Alps was the flight of E. Kloeckner, who on 11 October 1940 was the first one in history to reach the limit of the stratosphere in a glider. Figure 90 shows a schematic drawing of the waves utilized, on the basis of the observations of the flight and the indications recorded at the same time aboard a powered airplane.

The glider, type Kranich, was towed from its starting point at Ainring toward the west along the Alps, since a great wave cloud could be seen above the Wetterstein mountains north of Innsbruck. At an altitude of approximately 3000 m, at the level of the crest of the range, a violent turbulence inside a narrow layer of 200 m was found. At this time, a great wave cloud was developing over the Hohe Tauern range and the towing airplane turned east toward this cloud. The first slight downdraft was detected above Zell am See, and soon after that the first updraft. Between Zell am See and the Grossglockner, three waves were traversed in succession. The glider detached itself from the towing airplane at an altitude of 6480 m, northwest of the peak of the Grossglockner. It was then just ahead of a characteristic Foehn cloud. The vertical velocity was 1.5 m/sec with a horizontal velocity of 90 km/h (as compared to 67 km/h at 4000 m and only 18 km/h at 2000 m). South of the Alpine ranges, the cloud layer formed a Foehn wall as high as the Hohe Tauern range and disintegrated toward the north. The Foehn wave started directly above this wall of clouds. The cloud developed in a ~~northsouth~~ North-South direction over an expanse of 150 to 200 km, and in an East-West direction for at least 300 km. The glider climbed steadily at 15 m/sec and reached the base of the great wave at 8000 m, although the updraft had slackened between 7000 and 8000 m because of a momentary disintegration of the cloud. The pilot kept the airplane alongside the edge of the cloud until its contours had reformed again and the climbing speed was sufficient (2 m/sec). The cloud moved toward the south in the form of wedges, and reached a height of 14 km on the windward side. Figure 91 shows a similar cloud formation with such wedges. In this zone, at an altitude of 10000 m, Kloeckner found updrafts of 4 to 6 m/sec. However, the airplane was not winterized and the cold became hard to bear, even the airplane being affected by the low temperature. The pilot had to use both hands to move the stick, with the control mechanism making cracking noises at every movement and the fuselage creaking. At 11400 m Kloeckner decided to end the flight on account of the cold and exhaustion of his oxygen supply, even though the climbing speed was still 5 m/sec. For the descent he looked for a downdraft and lost altitude rapidly by flying tight spirals.

Thus a glider had for the first time reached the stratosphere, thereby proving that the Alpine Foehn wave propagates to very great altitudes. The principal Foehn wave reached an altitude of 14 km, and extended over 300 km, thus permitting long-distance flights similar to those carried out in cloud lanes by flying along the frontal zone of the Foehn cloud.

During Kloeckner's flight, a powered airplane flew above the great Foehn cloud between the Hohe Tauern range and the northern foothills of the Alps, in order to obtain data on the wave flow. The airplane was well-balanced, with constant speed and air intake, and the altitude variations recorded by the barograph proved definitely the existence of updrafts and downdrafts whose configuration suggested the shape of waves.

Analysis of the corresponding wave phenomena.

All these results are shown in Figure 90, in which elevations are shown greatly exaggerated.

At 4000 m altitude, stationary waves of small length (12 km) were registered. These were local waves to the leeward of isolated mountain peaks. They were created by isolated lenticules with a period of increase and decrease of about 15 minutes. Underneath these lenticules, between 2000 and 4000 m, the climbing speed reached 2 m/sec; above the cloud layer, the updraft stopped.

The flight path of the airplane showed a great wave (wavelength 40 km) at 7000 m altitude, beginning at the main range of the Hohe Tauern Mts. The second wave extended from the Leoganger and Loferer Steinberge to the foot of the Alps, also with a wavelength of 40 km. In this wave, in its lower part in the vicinity of the Leoganger Steinberge, a second edge with a typical lamellar shape could be distinguished. A third one, much less distinct, was located at the northern edge of the Alps and also had a wavelength of 40 km.

Scheme of the stationary waves above the Eastern Alps.

Other flights during Foehn periods were carried out over the Eastern Alps, between the Hohe Tauern Mts. and Lake Chiem. They confirmed the results of the flight of 11 October 1940, so that the characteristics of this wave system were definitely determined.

The Foehn over the Eastern Alps is split in two by a strong inversion layer as follows:

1) A lower flow, influenced by the ground and possessing local isolated waves. This layer reaches an altitude of 4000 to 5000 m.

2) A general wave flow, from 5000 m altitude to great altitudes. The temperature inversion which forms the retarding layer between these two flows is regularly produced during Foehn in the Eastern Alps and is generally located at about 4000 m altitude. In the lower flow, which is disturbed by terrain features such as peaks, mountain ranges, narrow or broad valleys, there is no organized wave flow, but local waves are formed behind the peaks or isolated mountain massifs and are characterized by lenticules of medium dimension. These local waves have a length of 10 to 15 km.

Above the inversion layer at 4000 m is the wave flow proper. Starting at 6000 m altitude, the full wind velocity is attained, which is never less than 50 km/h. In this upper flow, no longer influenced by local terrain features, but only by the massif of the Eastern Alps as a whole, the waves are shown by two great Foehn clouds, one between 6000 and 8000 m altitude and the other generally above 10000 m. The Foehn cloud extends over a considerable distance, 100 to 200 km, but it does not contain one uniform updraft field; rather, it is split by wave flow into updraft and downdraft fields, a fact which is frequently shown clearly by the appearance of its lower surface.

The lower cloud usually appears 30 to 40 km to the leeward of the principal mountain range, while the upper cloud begins directly above it and increases with the altitude in an upstream direction.

These statements show that it is possible to fly to very high altitudes - at least 15 km - in the general wave flow above an altitude of 4000 m in the Alps. At the present moment, howeve, it cannot yet be stated definitely that the upper wave can be reached directly from the ground.

Flights in the Innsbruck Region

This question is partially answered by the flights performed in the Innsbruck region, under the sponsorship of Dr. Hohenleitner. The starts were made halfway up a slope with an elastic rope, and the glider immediately made contact with the lower wave.

The location of Innsbruck is particularly typical and favors the formation of waves during a Foehn. About 30 km south of Innsbruck, the Brenner Pass (1370 m) makes a deep cut 10 to 20 km wide in the main mountain range whose peaks are all nearly 3500 m high.

The Foehn wall, which everywhere else remains on the windward side of the main chain, extends to the North half-way to Innsbruck, since the Wipp valley diverts the Foehn. The Foehn, once it has crossed the valley of the Inn, must climb over the steep Karwendel chain, which is 40 km long and 2500 m high. During the main phase of the Foehn, the flow from the South penetrates to the floor of the Inn valley, but is sometimes cut off by a thin layer of cold air. At any rate, it must climb over the wall of the Karwendel range and thus will produce a slope upwind of at least 2000 m altitude. The South-to-North cut shown in Figure 93, as established by air and ground observations, indicates that a stationary leeward wave is formed over the Inn valley, analogous to those observed over the Alps (Figure 94). This wave, often marked by a cloud, has been reached several times by gliders starting at an altitude of 900 m with an elastic rope. The gliders climbed to 2000 m over the slope in the north and then turned south across the valley to reach the wave. In this manner the vertical velocities prevailing in this region were determined. After leaving the slope upwind, the gliders first encountered a strong downdraft of minus 5 m/sec and then an updraft of plus 2 to 3 m/sec toward the center of the valley, and finally another downdraft of minus 5 m/sec on the south side of the valley. These vertical air movements permit us to fix the position of the wave right over the center of the valley.

One of these glider flights reached the lower portion of the great Foehn wave which is located at an altitude between 4000 and 6000 m. The pilot left the slope upwind at 2000 m, flew over the valley to the south and reached an updraft of 5 m/sec. He made contact with the front of the cloud of the upper wave at 3700 m and climbed along it to 4698 m altitude. At that altitude, the flight was broken off before reaching the upper limit of the cloud, but the pilot could see other very high wave clouds above it.

The local conditions at Innsbruck are shown schematically in Figure 94. Above 4000 m the wave system seems to correspond to that encountered over the Eastern Alps. The base of the great wave also starts above Innsbruck, 30 to 40 km from the main chain of the Alps. It is due to this identical location of the wave and the Karwendel mountains and of the local wave that contact between high-altitude waves and ground can be established.

Supplementary remarks on Alpine waves

a) Local waves

A few flights were carried out in the Bavarian Alps for the purpose of studying the isolated local waves which form in the lower portion of the flow affected by the terrain. These waves are found not only during a Foehn, but occur behind any isolated mountain massif properly situated and with wind velocities between 10 and 50 km/h. Figure 95 shows the barograms of two measuring flights in these local waves. In the first flight, the barogram shows that the vertical speed dropped. The pilot had missed a wave which was not distinguished by any cloud formation. In the second flight which followed immediately after the first, the wave was reached by a towed glider which then remained there in free flight for some time. In high mountains it is thus possible to fly continuously by means of these waves which are produced by slight winds, and to perform long-distance flights by using a technique similar to that of thermal flight, i.e., flying from wave to wave.

b) Characteristics of the later phases of the Foehn.

Until now we have dealt only with the main phase of the Foehn. In that phase, the upper wave with its well-defined edge is located directly above the main chain.

However, the end of a Foehn period is marked by the displacement of the wave front toward the northern edge of the Alps. The waves go beyond the Alpine ranges over the adjacent highlands. In October 1937 a flight was performed between Munich and Garmisch-Partenkirchen in the foothills of the Alps. Two distinct waves could be detected; the first, starting at the mountains, was particularly strong and had a length of 20 to 25 km. At 2800 m the variometer indicated a climbing speed of 6 to 7 m/sec. Between the two waves was a drowndraft which reached a velocity of 12 m/sec.

When the edge of the Foehn wave has finally reached a point above the Bavarian highlands, the impulse provided for the flow by terrain obstacles disappears. Thus the wave slackens and dies out.

7. Stationary stratospheric waves.

Deformations of the tropopause.

Kloeckner's flight proved indisputably that waves created by terrestrial obstacles can propagate to the stratosphere. Another proof can be furnished for this: The variation in the altitude of the surface of separation between troposphere and stratosphere. F. J. John has compared the temperature and pressure fluctuations recorded at Munich and at Pavia, on opposite sides of the Alps. He concluded from his measurements that the disturbance imparted to the flow by the Alps causes the variations in altitude of the tropopause to be much more frequent and more marked at Pavia and that the lower limit of the stratosphere is on the average lower at Pavia than it ought to be at that geographical latitude. Special examples demonstrate this variation in altitudes during a Foehn; thus, the measurements performed on August 19 and 20, 1925, show that, despite the influence of the latitude which normally lowers the tropopause, the tropopause is higher at Munich than at Pavia because of the forced oscillation over the Alps (Figure 97). This statement is of great importance for wave flying, to which the stratospheric regions are now being opened.

Observations pertaining to stratospheric waves.

The theoretical findings do not give us any indication as to how far up these waves extend, but we have been able to obtain some data from observations.

It has been known for a long time that certain clouds occasionally form what are called noctilucent clouds because of their irridescence. These clouds are particularly striking in South Norway. They were first described by Prof. Mohn of the University of Oslo. He noticed that these clouds, although at very high altitudes, are bound to certain periods and that they were observed at Oslo only when the Foehn was blowing. More recently, the altitude of these clouds has been determined by Prof. Stormer as between 20 and 30 km. He also stated that they appear during a Foehn and described their lenticular shape, which is exactly that of small common wave clouds. These observations are so clear that it can be assumed with great likelihood that these clouds are created by stationary Foehn waves (Figure 98), which can be reached by high-altitude gliders, as far as the sinking speed of the airplane, which increases with altitude, will permit; viz., the sinking speed of a normal glider which is 0.7 m/sec on the ground will be 1.8 m/sec at an altitude of 15 km; thus the updraft must be greater than this latter value.

Problems of stratosphere gliding.

Thus a new field of research has been opened. Its final developments cannot yet be foreseen, but its exploration has now become indispensable. It goes without saying that special techniques are required for stratosphere flights. The normal type of cockpit must guarantee a normal air supply and a certain temperature. For physiological reasons, the usual oxygen equipment is inadequate at altitudes above 12000 m. Therefore, a pressure cabin is necessary to ensure air supply and protection from the cold. There are no great dangers to be feared in stratosphere flight; nevertheless, provision must be made for bailing out at high altitudes, which requires the installation of special emergency equipment.

However, these special requirements are well within the capabilities of modern engineering and do not present any great obstacle to stratosphere flights.

8. Stationary waves over plains.

The existence of such waves.

Until now it had been believed that wave flight was possible only in hilly regions, because of the disturbance of the general flow by the terrain. According to observations made so far, the best conditions combine the occurrence of a Foehn wind with a location in high mountain terrain.

However, the Foehn is a local phenomenon and thus limits wave flight to certain localities, just as slope updrafts limit the glider to flight over slopes, except in special cases. It is therefore natural that the question should arise whether the region of wave flight could not be extended if stationary waves, independent of mountain massifs, could be found over plains.

The basic wave disturbances are not started by fixed obstacles such as mountains, but may be initiated by some kind of a modification of the lower flow. The progression of the systems may be accompanied by variations in wind velocity which will influence the flow at higher levels in the same manner as fixed obstacles, and will thus create wave fronts which are stationary with respect to the system in motion. A layer of cold air is like a mountain, so that one is led to believe that it would create a stationary wave system.

A. Defant has investigated those disturbances of the lower troposphere, such as cold fronts, which cause variations in the level of the tropopause. His investigations were strictly from the meteorological point of view. He concluded that any kind of disturbance of the lower troposphere, caused either by variations in horizontal velocities or by breakthrough of cold air masses underneath the general West flow, will cause wave-like variations in the level of the tropopause.

Thus we can expect to find changes in the level of the tropopause and wave-like disturbances in it, in front of a mass of moving cold air, analogous to the disturbances set up by a mountain obstacle. These waves are stationary with respect to the cold front and should be accompanied by an updraft field similar to that of the Foehn waves. The assumed structure of these waves is shown in Figure 99.

The typical lenticular clouds which are forerunners of bad weather can frequently be seen in the still clear sky. Over the plains, especially near the seacoast, these clouds are most frequently seen, undoubtedly because the atmosphere there is not disturbed.

Explanation of front flights.

The existence of these waves is frequently confirmed by glider flights. The barograms shown in Figures 64 and 65 are particularly significant from this point of view.

In the barogram of Figure 64, the glider remained ahead of the front at an altitude of 2900 m for about 50 minutes, after a rapid climb. The flight continued after a loss of 400 m of altitude, with a strikingly regular succession of climbs and descents for over two hours. Such a barogram is obviously different from those obtained in normal thermal flight or in turbulences preceding cold fronts. It shows a stationary updraft field ahead of the front and also a wave-like structure of the field.

The two barograms taken on 25 July 1931 (Figure 65) confirm the existence of this regular stationary updraft field ahead of a front. After reaching the updraft in front of the thunderstorm, the two gliders climbed rapidly to 2500 m and then remained at altitudes between 2300 and 2500 m. One of the barograms has a peak which corresponds to the terrain, because the glider in question was flying above the Thuringian Forest at that time. This constant and rapid climb in front of the thunderstorm and outside the cloud zone, which took place in the same manner for both gliders, although time and place were different, bears a striking resemblance to the climb in a Foehn wave.

Since that time front flights have been rare; this theory came up only as an afterthought, so that no definite proof is available. However, the resemblance of the various barograms of front flights is such that a few experimental flights will suffice to settle the issue. These wave flights ahead of a front are neither difficult nor dangerous. It is possible to tow the glider to a point ahead of the front and then, in free flight, the pilot must merely see to it that he does not fly into the front itself. The solution of this problem will be a new and important step forward in gliding, since it will be the first attempt to make wave flight independent of mountains.

Other types of waves over plains.

There are certainly other possibilities of wave flight in plains. Huge wave clouds have frequently been observed at altitudes between 6000 and 7000 m; a particularly striking specimen appeared over Paris on 17 March 1947. These waves look exactly like the high Foehn waves and strengthen the conviction that wave flight over plains is possible.

In support of this theory, a measuring flight carried out in a Junkers A 35 airplane can be cited. At 4400 m altitude, with constant deflection of the elevator, the climbing speed varied greatly during the flight underneath an altocumulus formation, going from 1 m/sec to 3.1 m/sec, the wave-length being of the order of 4 km.

All sorts of disturbances can cause the formation of waves over flath terrain. Great differences in surface roughness, such as transitions from water to land surface or great changes in wind direction near the ground, suffice.

Thus a great new field of research has opened for glider flight, which will have to be illuminated step by step in the course of the next few years.

- END OF CHAPTER IV -

CONCLUSION

Throughout its twenty-five years of existence, glider flying has achieved performances which never could have been predicted in advance, and this has been done through sheer effort and despite its simple and often rediculed beginnings.

We have now reached a turning point in its development. The troposphere -- that is, the first ten kilometers of our atmosphere -- has been conquered, and the first penetration into the stratosphere has been successful. The methods by which glider flying can reach these altitudes have been worked out; and for those who know the enthusiasm and unselfishness of glider pilots, there is no doubt that the problems which are still unsolved will be solved in the near future.

The secret of success lies in a close collaboration between theoretical research and practical glider flying. Science has opened up new roads to glider flying, but the flights themselves have opened up perspectives for science. Flight performances have greatly assisted in increasing our knowledge of the free atmosphere. Aerology and meteorology are particularly indebted to glider flying for much fundamental data. The glider has become the best instrument for measuring vertical air movements and, in contrast with the other measurement procedures, it furnishes quantitative results.

Although the physics of Foehn has long since been studied in all its details, it is glider flying which discovered the great foehn wave, an essential addition to the fundamentals of Foehn dynamics.

In view of all this, one can well wonder why the glider has not yet become a regular instrument of research, with all the possibilities it offers to both aerology and aerodynamics.

The pilot who has followed the author throughout this book can understand the motives underlying the three qualities demanded of a glider pilot: a mastery of the art of piloting, scientific curiosity regarding the problems of glider flying, and enthusiasm for its beauty. These qualities instill in the pilot the ideal necessary to the success of his flights, which unselfishly serve both the progress of glider flying and human knowledge.

Glider flying has had the singular good fortune of not becoming enslaved to an objective. May it keep this gift as the basis of its performance and its spirit.

(List of 99 Figures appended, with their captions.)

- E N D -

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FIGURES WITH THEIR TITLES

LIST OF ~~ILLUSTRATIONS~~

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Epaisseur de la couche limite - depth of boundary layer

P. 5 on microfilm

Descente - descent

Montée - climb

Figure 2 - a) Flow lines along a ridge

P.5 on microfilm

b) Velocity distribution in the horizontal plane

Figure 3 - Flow over a ridge

P.7 on microfilm

a) Potential flow

b) Formation of eddies

c) Row of eddies and dead-air zone

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Inconnu - unknown

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Figure 10 - Traces of smoke clouds

Dune - dune

baie - bay

vent - wind

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Vent - wind

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vent -wind

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Couche limite - boundary layer

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b) Flow during the night. The valley is filled with cold humid air.

Air froid immobile - immobile cold air. Use photostat from here

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T - adiabatic of still air

T'- adiabatic of an ascending or descending air particle

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Figure 20 - Unstable equilibrium p 17 ~~on microfilm~~

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Altitude de l'équilibre thermique - altitude of thermal equilibrium

Figure 22. No title p 20 ~~on microfilm~~

Altitude de l'équilibre de .. - altitude of equilibrium of ..

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Coucher du soleil - sundown

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Figure 26 - Recording of the pulsations of the wind direction and velocity on the ground

p 22 ~~on microfilm~~

Figure 27 - Measuring of a miniature whirlwind by a meteorological balloon.

Echelle - scale p 23 ~~on microfilm~~

laché a bord d'un avion - release from airplane

Niveau de la mer - sea level

Vitesse du vent - wind velocity

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Echelle - scale p 24 ~~on microfilm~~

Altitude en m au-dessus du niv, de la mer - altitude in meters above sea level

V en m/sec - velocity in m/sec

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lache à bord d'un avion - release from airplane p 25

Altitude en m au-dessus du niveau de la mer - Altitude in m above sea level

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Altitude en m au-dessus du niveau de la mer - Altitude in m above sea level

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Figure 37 - Diagram of temperatures during land and sea breezes at Rio de Janeiro.

Aerological sounding performed by airplane. p 28

Adiabatique humide - humid adiabatic

Brise de mer - sea breeze

Brise de terre - land breeze

Figure 38 - Temperature curves for East and West winds, at Homs, Libya.

Vent d'Est - East wind

Vent d'Ouest - West wind

Figure 39 - Diagram of the temperature on 2 April 1939, at Homs, Libya, and the maximum altitude reached by the glider.

Plafond - ceiling

Figure 40 - Curve of mean ~~xxxx~~ temperature at Garian, Libya, 9 and 17 April 1939, at 1300 hrs.

Figure 41 - Curve of temperature at Garian, during a West wind. Maximum altitude reached by glider.

p 29

Figure 42 - Barogram of a glider flight from Homs to Buenat-el-Hsun, showing the effect of ~~the~~ land and sea breezes on the altitude of the glider.

p 30 on

Région soumise à l'influence de la brise de mer - Region under the influence of the sea breeze

Région soumise à l'influence de la brise de terre - Region under the influence of the land breeze.

Figure 43 - Temperature curve in unstable equilibrium p 31

Altitude de l'équilibre - equilibrium altitude

Altitude de condensation - Condensation altitude

Figure 44.- Unstable-humid atmosphere; upward velocity of an isolated mass of humid air. T- temperature of the surround air vertically at rest. T'-dry adiabatic. T"- humid adiabatic. C - curve of vertical velocity. p 32

Adiabatique humide - humid adiabatic

Adiabatique sêche - dry adiabatic

Vitesse verticale - vertical velocity

Figure 45 - Flight path of a glider in a cumulus cloud. Velocity of wind at ground level 11 m/sec. Sky 9/10 overcast (cumulus). p 33

Temps en min.- time in minutes

Figure 46 - Flight path of a glider in a cumulus cloud. p 34

Altitude en m - altitude in m

Temps en min. - time in minutes

D'après le vol plané - Velocity after downward glide

Variation de temp. à partir de 14h10 - Temperature variation after 1410 hrs.

Tempér. en °C - Temperature in °C

Figure 47 - Inversion condition. p 35

Altitude en m - altitude in m

Tempér. en °C - temperature in °C

Pression en mm - pressure in mm

Figure 49 - Map of long-distance flights carried out in Libya, 1939 p 36

Mediterranée - Mediterranean

Figure 50 - Diagram obtained with a recording device p 37

Pression dynamique - dynamic pressure

Pression atmosphérique - atmospheric pressure

Accélération verticale - Vertical acceleration

Figure 51 - Speed polar of the "Wanderkranich" glider

Figure 52 - Schematic drawing of a measuring flight

Descendance - Downdraft

Ascendance - Updraft

Vue en plan - horizontal section

p 38

Figure 52 - cont'd

Trajectoire de l'avion - Flight path of the airplane

Vol rectiligne pendant les mesures - Rectilinear flight during measurement

Vol en virage sans mesure - Turns, no measurement

Figure 53 - Field of vertical winds underneath flattened cumuli

Isotherme - isotherm p 38

Vent - wind

Nuages:2/10 cumulus, le 3 Août 1938, 13h44 ~~xxx~~ à 14h47.

Sky:2/10 overcast (cumulus). 3 August 1938, 1344 hrs to 1447 hrs.

Altitude en m - altitude in m

Figure 54 - Field of vertical wind underneath a cumulus. p 39

Wind: WNW ; 1 to 4 m/sec; temperature gradient per 100 m:0.85°C up to 2500 m. Sky 8/10 overcast (cirrus and cumulus). 20 July 1937.

Direction du vent - Wind direction. à - to

Figure 55 - Thermal exchange movements.Wind: SE, very slight. Sky; 2/10 overcast (flattened cumulus). 23 June 1937, 1415 hrs to 1535 hrs. p 41

à -to

Figure 56 - Eddies in parallel rows, after Idrac. p 42

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Figure 58 - Formation of clouds in parallel rows. p 42

a) Horizontal section. b) Vertical section. c) Flow underneath cumuli arranged in rows.

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b) Schematic representation of the flow of air around Mt.Cervin, ~~xxxxxxx~~ after Kampé de Feriet. p 44

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Figure 62 - Schematic drawing of thermal gliding and of cloud lane gliding.

Thermiques normaux - normal thermal currents p 45

Rues de nuages - cloud lane

Trajectoire du vol - flight path

Direction du vent - wind direction

Barogramme - barogram

Figure 63 - Schematic drawing of a cold front p 45

chaud - warm

froid - cold

Figure 64 - Barogram of the first front flight, July 1929 p 46

Figure 65 - Barogram of two front flights p 46

Figure 66 - Temperature-altitude diagram during thunderstorm conditions p 48

Altitude en m au-dessus du niv. de la mer - Altitude in metres above sea level

Temp.de l'air du nuage - temperature of the air ~~of~~ in the cloud

Temp.de l'air environnant - temperature of the surrounding air

Temp. de l'air sec - temperature of the dry air

Adiabatique humide - humid adiabatic

Adiabat. sèche - dry adiabatic

Vitesse de montée - climbing speed

Figure 67 - Barogram of an altitude flight in a thunderhead p 49

Figure 68 - Structure of a thunderstorm cloud p 50

Neige - snow

Glace - ice

Grèle - hail

Eau - water

Zone de givrage - icing zone

Nuage de grain - squall cloud

Pluie - rain

Chaud - warm

Froid - cold

à -to

Pseudo-front froid - cold pseudo-front

Isotherme - isotherm

charges electriques - electric charges

Figure 69 - Flow lines of a wave p 53

Figure 70 - Recording of atmospheric waves from an airplane p 53

Humidité - humidity

Température - temperature

Figure 70 - cont'd

Vitesse - speed

Régime moteur - Operation of engine

Pression statique - static pressure

Figure 71 - Path of waves measured by meteorological balloons, 2 April 1930,1620 hrs.

Altitude ... - Altitude above sea level, in meters p 54

Figure 72 - Path of waves measured by meteorological balloons, 2 April 1930, 1200 hrs.

Altitude .... - Altitude in meters above sea level p 55

Figure 73 - Barogram of a long-distance flight under the influence of waves

Temps de vol en minutes - Flight time in minutes p 55

Figure 74 - Formation of clouds of free waves p 56

Bancs de nuages - cloud banks

Vent inférieur - lower wind

Vent supérieur - upper wind

Figure 74 - Formation of waves on the free surface of a liquid p 57

a) Flow velocity less than velocity of propagation of the wave

b) Flow velocity greater than velocity of propagation of the wave

c) Transition from "stream flow" to "torrential flow"

Figure 76 - Formation of waves on the free surface of a liquid p 57

a) Flow velocity less than the velocity of propagation of the wave

b) Flow velocity greater than the velocity of propagation of the wave.

Figure 77a- Distribution of upwinds below an obstacle p 59

Figure 77b- Distribution of upwinds of waves over a plateau p 59

Figure 78 - Flight path of a statically balanced balloon to the leeward of a mountain massif p 60

en minutes - in minutes

Figure 79a- Flight paths of statically balanced balloons;wind velocity 11 m/sec.

Note: The elevations are exaggerated ten times.

Ballon - balloon. p 60

Figure 79b- Flight paths of statically balanced balloons behind dunes.

Distances au point de départ - distances from starting point

Ballon - balloon

Altitude en m - altitude in meters p 60

Figure 80 - a and b: Flight paths of statically balanced balloons showing stationary waves. p 61

Figure 81 - Flight path of statically balanced balloons to the leeward of a mountain massif

en - in p62

Figure 82 - Schematic drawing of Foehn waves to the leeward of the Giant Mountains

Trou de Föhn - Foehn pocket

1ère, 2ème, 3e onde - 1st, 2nd, 3rd wave

tourniquets - whirls p 63

Moazagotl - "Moazagotl" Foehn cloud

Figure 83 - Distribution of updraft and downdraft fields downstream from the Giant Mts.

zone ascendante - updraft zone

zone descendante - downdraft zone p 64

0 à 500 m - 0 to 500 m

au-dessus de - above

Coupe - section

Figure 84 - Cloud of a high-altitude wave above the Giant Mountains Photographs - not reproducible. p 65

Figure 85 - Wave clouds in the lee of the Cevennes Mts.

Figure 86 - Roll of eddies

Figure 87 - Barogram of a wave flight over the Giant Mountains p 66

Départ aérodrome de Hirschberg 8h12 - Start at Hirschberg airport 0812 hrs

Atterrissage aeroport de Breslau 15h54 - Landing at Breslau airport 1554 hrs

Largage - release from towing airplane

Figure 88 - Wave cloud above the Alps Photograph - not reproducible. p 66

Figure 89 - Field of updrafts over the Eastern Alps, computed after Lyra's formulas

Vent - wind p 67

Figure 90 - Wave system over the Eastern Alps, after measurements performed simultaneously by glider and by powered airplane p 68

Vol remorqué - towed flight

Vol à voile - gliding

Trajectoire de l'avion - flight path of the powered airplane

Plafond - ceiling

1ère, 2ème, 3ème onde -1st, 2nd, 3rd wave

Fig. 90 -cont'd
Grande onde de Fohn(Moazagotl) - Great Foehn wave ("Moazagotl")

Sud - South

Altitude de largage - Release altitude

Vent - wind

Lenticulaires isolées -Isolated lenticular clouds

Nuage de refoulement du Föhn - cloud of slowed-down Foehn wind

altitude en km - altitude in km

Figure 91 - Leading edge of a Foehn wave Photograph - not reproducible. p69

Figure 92 - Map of the Innsbruck region p 70

Crête du Nord - North Chain (Nordkette)

Vallée de l'Inn - Inn Valley

Col du Brenner - Brenner Pass

Figure 93 - Schematic drawing of the Foehn waves over the Alps near Innsbruck. p 71
I. Local waves, wavelength 10 to 15 km and altitude 2 to 4.5 km. II. Lower great Foehn waves, wavelength 40 to 45 km and altitude 6 to 8 km.
III. Upper great Foehn waves, wavelength 40 to 45 km and altitude above 9 km.

Vent horizontal > 50 km/h - horizontal wind above 50 km/h

Col du Brenner - Brenner Pass

Sud - South

Nord - North

Altitude en km - altitude in km

Figure 94 - Schematic drawing of flight underneath waves p 72

Nord - North

Vent ascendant de pente - slope upwind

Crête septentrionale - North crest

Vent ascendant des ondes locales - Updraft of local waves

Crête méridionale - South crest

Vent descendant - downdraft

Sud - South

Figure 95 - Barogram of flights under the influence of local waves p 73

1er vol, 2ème vol - 1st, 2nd flight

Figure 96 - Wave clouds over Upper Bavaria p 73

Vent - wind

Vallée de Ziller - Ziller valley

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Figure 97 - Measurements of temperature above Munich and Paris, and altitude of the tropopause (taken on ~~xx~~ 19 and 20 August 1925). p 74

Température en degrés absolus - Temperature in °K.

Figure 98 - Noctilucent ~~Luminescent~~ clouds p 75 Photograph - not reproducible.

Figure 99 - Schematic drawing of waves ahead of a cold front p 76

Tropopause - tropopause

Inversion - inversion

Air froid - cold air.

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OF LIST OF
FIGURES

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